JOURNAL OF ATMOSPHERIC AND TERRESTRIAL PHYSICS

EDITORIAL ADVISORY BOARD

H. O. G. ALFVÉN–STOCKHOLM · **J. BARTELS**–GÖTTINGEN · **P. M. S. BLACKETT**–LONDON · **E. G. BOWEN**–SYDNEY · **SIR DAVID BRUNT**–LONDON · **R. BUREAU**–BAGNEUX · **S. CHAPMAN**–ALASKA · **J. EGEDAL**–COPENHAGEN · **H. FICKER**–VIENNA **V. F. HESS**–NEW YORK · **E. O. HULBURT**–WASHINGTON, D.C. · **H. ISRAEL**–AACHEN **D. K. MENZEL**–CAMBRIDGE, MASS · **S. K. MITRA**–CALCUTTA · **M. NICOLET**–UCCLE **E. H. NORINDER**–UPSALA · **B. van der POL**–GENEVA · **N. V. PUSHKOV**–MOSCOW **J. A. RATCLIFFE**–CAMBRIDGE · **O. E. H. RYDBECK**–GOTHENBURG · **SIR GEORGE SIMPSON**–LONDON · **H. SOLBERG**–OSLO · **J. VELDKAMP**–DE BILT · **M. WALDMEIER**–ZÜRICH · **J. ZENNECK**–ALTHEGNENBERG

EDITOR IN CHIEF

SIR EDWARD APPLETON, F.R.S.

EDINBURGH

A SPECIAL SUPPLEMENT

1957

PROCEEDINGS OF THE POLAR ATMOSPHERE SYMPOSIUM

PART I

Meteorology Section

Held at Oslo, 2–8 July 1956

PERGAMON PRESS

NEW YORK · LONDON · PARIS · LOS ANGELES

Sup
Par
195

Suppl.
Part.I
1956

CONTENTS

PAGE

OPENING ADDRESS - - - - - - - - - vii
F. L. Wattendorf

INTRODUCTION SPEECH - - - - - - - - - ix
H. U. Sverdrup

SECTION I

1. ARCTIC GEOGRAPHY AND CLIMATE - - - - - - 3
F. Kenneth Hare

2. SIMILARITIES AND CONTRASTS BETWEEN THE ARCTIC AND ANTARCTIC MARINE CLIMATES - - - - - - 9
R. M. Holcombe

3. THE RECENT VARIATIONS OF THE CLIMATE AT THE NORWEGIAN ARCTIC STATIONS - - - - - - - - - 18
Th. Hesselberg and T. Werner Johannessen

4. A BRIEF REVIEW OF RESEARCH IN ARCTIC METEOROLOGY - 30
A. D. Belmont

DISCUSSION, 2 JULY 1956 - - - - - - - - 37

SECTION II

1. THE GENERAL CIRCULATION OF THE LOWER TROPOSPHERE OVER ARCTIC REGIONS AND ITS RELATION TO THE CIRCULATION ELSEWHERE - - - - - - - - - 45
Jerome Namias

2. RECENT STUDIES ON THE ARCTIC TROPOSPHERE AND ITS TELECONNECTIONS - - - - - - - - - 62
H. Flohn and G. Seidel

3. WINDS AND TEMPERATURES IN THE ARCTIC STRATOSPHERE - 71
H. Wexler and W. B. Moreland

4. THE ARCTIC WINTER STRATOSPHERIC JET STREAM (ABSTRACT) 85
W. L. Godson

5. THE PROBLEM OF THE MAINTENANCE OF THE LOW-LEVEL ANTICYCLONIC CIRCULATION IN THE ARCTIC - - - 87
R. Fjørtoft

DISCUSSION, 3 JULY 1956 - - - - - - - - 93

SECTION III

1. THE ROLE OF TROPOSPHERIC COLD-AIR POLES AND OF STRATOSPHERIC HIGH-PRESSURE CENTERS IN THE ARCTIC WEATHER 101
R. Scherhag

PAGE

2. Arctic Weather-forecasting Problems of the Air Weather Service - - - - - - - - 118
R. D. Fletcher

3. Arctic Weather Analysis - - - - - - - - 124
R. J. Reed

4. The Quantitative Representation of the North Polar Pressure Field - - - - - - - - - 137
F. Kenneth Hare

5. Discussion, 4 July 1956 - - - - - - - - 151

6. An Empirical Method of Forecasting 24-hour Movement and Change of Intensity of Sub-Arctic North Atlantic Cyclones (Abstract) - - - - - - - - 156
J. J. George, P. M. Wolff and W. L. Somervell, Jr.

SECTION IV

1. Solar Radiation Measurements in the Arctic Ocean - 159
S. Fritz

2. Long-wave Radiation and Turbulent Heat Transfer in the Antarctic Winter and the Development of Surface Inversions - - - - - - - - - - 167
G. H. Liljequist

3. On the "Arctic Whiteout" - - - - - - - - 182
S. Fritz

4. On the Relation between Meteorological Conditions and Total Amount of Ozone over Tromsö - - - 187
H. Johansen

5. Visual Range in the Polar Regions with Particular Reference to the Alaskan Arctic - - - - - 195
J. Murray Mitchell, Jr.

6. Some Problems of Meteorological Observing in Polar Regions - - - - - - - - - - 212
N. J. Schumacher

7. Lower Tropospheric Inversions at Ice Island T-3 - - 215
A. D. Belmont

8. Current Status of Sea Ice Reconnaissance and Forecasting for the American Arctic - - - - - - 285
C. C. Bates

9. Meteorological Conditions and the Associated Sea Ice Distribution in the Chuckchi Sea during the Summer of 1955 - - - - - - - - - - - 323
J. W. Winchester and C. C. Bates

Discussion, 6 July 1956 - - - - - - - - 335

Suppl.
Part.I
1956

OPENING ADDRESS

by

FRANK L. WATTENDORF

Director

ADVISORY GROUP FOR AERONAUTICAL RESEARCH AND DEVELOPMENT

of the

North Atlantic Treaty Organization

ON behalf of AGARD, it is a great pleasure to welcome the participants in this Symposium on the Polar Atmosphere and to extend a special greeting in the name of our distinguished chairman, Dr. Theodore von Kármán, who is presently in California.

Many of you may not be too well acquainted with AGARD, since this is the first symposium we have held in this particular field. I think therefore it might be appropriate to review some background information on our organization, the Advisory Group for Aeronautical Research and Development (AGARD).

AGARD is really the brainchild of Dr. von Kármán. From his earliest years he was influenced by the thinking of his father, Professor Maurice von Kármán, a prominent philosopher with constructive international views. This gentleman, a professor of philosophy in the University of Budapest, Hungary, as early as 1900 expressed the opinion that in about fifty years there would arise organizations for active international co-operation; that the chief hope of lasting peace was for all nations to work together for the common good.

It is interesting to note that it was just about fifty years later that NATO came into being, and that his son, Dr. Theodore von Kármán, used that as a framework to establish an international organization of scientific co-operation.

The more recent history of AGARD dates from 1945 when Dr. von Kármán was called upon by General H. H. Arnold, Chief of the United States Army Air Forces, to advise him on the shape of things to come.

As an initial step, Dr. von Kármán gathered together a small group of scientific and technical experts, and with them visited Europe in April 1945. After this visit he was more than ever convinced that the most important thing for the future was that all nations should work together in scientific and technical fields. To further this ideal, he recommended the formation of an international Scientific Advisory Board. Several years were to pass before there was a mechanism to form such an organization.

Finally, when NATO came into being, it provided a potential framework. Dr. von Kármán, with the support of the United States Air Force, called a preliminary meeting in 1951, in Washington, of Research Directors of the NATO countries. This meeting unanimously recommended the formation

of an Advisory Group for Aeronautical Research and Development to the North Atlantic Treaty Organization. This was approved in February 1952 as an organization reporting directly to the Standing Group; in other words on the same organization level as SHAPE.

The objectives of AGARD are to recommend ways and means of increasing the NATO research and development potential. The approach is by taking limited constructive steps. Instead of trying to cover the whole field, AGARD acts as a catalyst and stimulates work in certain fields where there are known deficiencies. In particular, several technical panels were formed in fields where tangible accomplishments could be expected in relatively short time. Then there was established an exchange program, whereby young scientists and engineers of different nations were given an opportunity for training and experience in laboratories of other nations. From time to time, special committees are set up to advise on specific problems referred to AGARD by SHAPE and NATO Agencies. Finally, there are certain fields where it has been found more expedient to hold exploratory symposia devoted to topics not wholly covered by other organizations. The present symposium comes under the latter category.

At the Second AGARD General Assembly in Rome in December 1952, the delegates of the different NATO nations were asked to make recommendations as to fields of activities which would be of particular importance to their own nation. The representative of Norway, Dr. Harang, who is with us today, stressed the importance of the polar atmosphere, and recommended AGARD activities along this line. It was decided that the way to approach the project was to hold an exploratory symposium assembling the appropriate experts of different NATO nations to examine the more pressing problems of the polar atmosphere; then conclude from their findings the best way of proceeding in the future. Out of the above considerations came the recommendations for this particular symposium.

Since Norway had made the original suggestion, it was appropriate that Norway be invited to plan the program. The Executive Committee for this symposium is under the chairmanship of Professor Harald Sverdrup* and the Arrangement Committee is headed by Dr. Harang.

It is interesting to note that the polar atmosphere is of prime importance to both commercial and military aviation, and to advancing knowledge in the geophysical sciences, including weather prediction and control. For this reason, this symposium is quite in spirit with the new look in NATO; in other words, the placing of greater emphasis on Article 2 of the North Atlantic Treaty, which concerns activities contributing toward the further development of peaceful and friendly international relations. A committee of three foreign ministers was appointed to provide proper recommendations. It is notable that among the "three wise men" chosen to make this study, the Foreign Minister of Norway is one. It is also significant that this Polar Symposium was already in preparation well in advance of the Council Meeting, which is a tribute to the vision and forward thinking of the scientists of Norway in this regard.

In conclusion, I wish to extend the personal best wishes of our chairman and the AGARD organization for a successful meeting.

* Deceased August 1957.

Supp
Part
1956

INTRODUCTION SPEECH

H. U. SVERDRUP*
Norsk Polarinstitutt, Oslo, Norway

Dr. Wattendorf, Delegates and Guests,

IT is a great pleasure to me, on behalf of the Executive Committee from Norway, to welcome all the delegates and guests to this meeting.

I dare say, when it was first suggested that such a meeting should be held here in Oslo, we who were asked to take steps towards arranging it, hesitated quite a great deal for two reasons. On the one hand, Norway lies in a corner of Europe, somewhat out of the way, although, thanks to the air communications, distances have been reduced so much that after all the location is not all-important.

Of greater concern was that we might not be able from here to contribute so much to the meeting as we would like to do. But when we still went ahead, the reason was that we felt that a meeting of this sort would be a tremendous stimulus to the work going on in this country.

The particular reason why I am standing here as chairman of the Executive Committee of Norway is perhaps that I am a sort of exhibit A, because it so happens that I am the oldest living meteorologist who has worked actively in the Arctic. I cannot claim the same position when it comes to the Antarctic, where Sir George Simpson, who is still going strong, took part in the British Antarctic Expedition 1910–13, quite a number of years before I went into the Arctic. There is also another reason why I may be exhibited here, namely that I think that of all present I have spent more time in the Arctic than anybody else, and I also believe that you had to get together several of those present here, who worked in the Arctic, in order that their combined time of services would equal to the one that I have behind me. Therefore, to me this meeting here is an historical event, because it will indeed throw light over the tremendous development which has taken place during the 38 years now which have passed since in 1918 I left northern Norway on board Roald Amundsen's ship the *Maud*. The changes which have taken place cover not only the means of transportation and communications, but also the instrumental equipment and the possibilities of undertaking systematic research in the very high latitude. As far as communications are concerned I may mention that, when we left in 1918, we were prepared upon being absent for four years and carried no radio. Nothing would be heard about the expedition until it returned, if it returned. Actually, however, it so happened that the expedition fell in two parts, and it took seven and a half years before it was back again in the country.

But it is not only the communications, it is particularly the types of observations which have been subject to enormous changes. Going back

* Professor Sverdrup died suddenly in Oslo on 21 August 1957.

again, the only work pertaining to the ionosphere, which was done at that time, was the study of the aurora. The technique of measuring the height of the aurora had been developed, and the analysis of the spectrum of the aurora had been started, but the entire knowledge of the ionosphere was based upon the indeed difficult interpretation of the results of the aurora observations. Turning to the atmosphere, the examination of the free atmosphere had to be undertaken exclusively by means of pilot-balloons and kites. By means of the latter one could not reach appreciable altitudes in spite of the often back-breaking work of hauling down the kite against a strong wind. But the expedition to which I refer was actually the last of the long-lasting pioneering expeditions to the Arctic. I have always looked upon the period of these expeditions as a period in which two things were accomplished. In the first place the geographical exploration was carried through, not completed, but to a considerable degree brought to an end. In the second place these expeditions demonstrated fully the possibilities of survival in the Arctic, and that under the rigorous conditions of the Arctic one could use the same types of instrumental equipment which we employ in lower latitudes. From these points of view the pioneering attacks have been amply rewarded, because now we find permanent stations established at the most remote localities, and we find them equipped with the most complete instrumentations.

The changes which have taken place are particularly reflected in the titles of the papers to be presented here now at this symposium. Within meteorology, for instance, most of the papers deal not only with surface conditions, but partly or exclusively with conditions of the free atmosphere. Within the ionospheric section, about which I know very, very much less, practically nothing, the situation is indeed that there is not a single paper dealing with the aurora, but, instead, there are papers there which are of direct interest to the meteorologists, dealing with the drifts and turbulence in the ionosphere. This overlapping of interests now between the meteorologists and the group working with the ionospheric problems is paid a certain attention to in arranging this programme, because on Thursday there will be no session of the meteorological section, but the ionospheric section will then deal with papers which are of interest also to the meteorologists.

A symposium on the polar atmosphere seems indeed justified from the point of view that it is desirable to review the recently gathered knowledge and to exchange views as to the interpretation of the results. However, when this symposium was first discussed, an objection was raised: why a special symposium on the polar atmosphere? Can we deal with the polar atmosphere without considering that it is not isolated, without considering that the atmosphere of the northern region plays a role in the general circulation similar to that of the atmosphere above any other part of the world? I am now referring particularly to the meteorological part of the programme. Obviously we cannot neglect the importance of the characteristics of the polar atmosphere to the general circulation, and this topic will therefore also be dealt with here, but still I believe that it is quite justified to talk about polar meteorology as a special subdivision within meteorology. We speak of tropical meteorology, and it seems justified to talk about the

meteorology of the middle latitudes where conditions both from the tropical and from the polar parts of the atmosphere will have to be taken into account. However, it is not only because of certain climatic contrast that one conveniently introduces the subdivisions "tropical" and "polar" meteorology. There are certain essential differences. When dealing with the dynamics of the systems, there are differences because of the latitude change of the Coriolis-acceleration, but fundamentally the differences within the dynamics are not especially important. The fundamental difference is found when one considers the energy transformations within the atmosphere in these two regions. In the tropics the most important energy transformations are represented by the addition of latent heat to the atmosphere through the condensation of water vapour. That, besides the phenomena associated with radiation, leads to instabilities which play a tremendous part in all problems pertaining to the meteorology of the tropics. In the polar regions, on the other hand, the most important energy transformations are related to processes of radiation, and a characteristic of the polar atmosphere is that processes of instability are there of very much smaller importance. This also is reflected in the titles of papers to be presented here, because there is not a single paper which deals with questions pertaining to precipitation or to cloud-formations in the polar atmosphere, whereas there are several papers dealing with processes of radiation and dealing with the types of inversions which are typical under stable conditions. Because of the character of the energy transformations I considered that it is fully justified to speak about polar meteorology in contrast to tropical meteorology.

In concluding this very brief introduction I would like to return to the character of this symposium. I believe that it will serve particularly two purposes. It will help to clarify a number of the problems of the polar regions and, in the second place, it will stress the importance of developments in the polar regions to the problems in our latitudes. In that respect it will of course deal not only with the strictly polar conditions, but also with those which concern the regions around the Arctic with their very dense population and with their increasingly important problems in meteorology.

We who have had to deal with the arrangement of the meeting are pleasantly surprised by the attendance which the meeting has obtained and by the fact that so many contributions will be presented here. We are also, I would add, extremely grateful to the different speakers because they have conscientiously sent in abstracts of their talks at such an early date that the abstracts are available now. We feel that what little effort we have been able to put into the arrangement has indeed been amply rewarded by the attendance, and by the interest shown in the meeting which starts here now. In view of that background I feel confident that some of the goals of the meeting here, which were also mentioned by Dr. Wattendorf, will be attained.

Sup
Par
195

SECTION I

Suppl.
Part.I
1956

Sup
Par
195

ARCTIC GEOGRAPHY AND CLIMATE

F. KENNETH HARE
McGill University

The Arctic and Sub-Arctic cannot readily be defined in terms of atmospheric circulation. Instead one must depend on certain geographical distributions in partial equilibrium with the existing climates. These include (1) the zonation of natural vegetation, (2) the occurrence of permafrost and patterned ground, (3) the distribution of sea, lake and river ice, (4) the continental glaciers, and (5) the circulation of Arctic and Sub-Arctic water masses. The mechanism of climate control over these distributions can be revealed only by physical meteorological research, at present rarely undertaken. A map is presented showing the basic distributions as at present known.

Suppl.
Part. I
1956

INTRODUCTION

THE broad, rather ambiguous title of this paper was suggested by the organizers of the symposium, and not by the author, who takes a more modest view of his ability to cover so broad a field. He was asked to present, in the opening paper of the symposium, a review of the way in which the northern atmosphere interacts with other aspects of the physical geography of the Arctic and Sub-Arctic. The author does not know exactly what was in the minds of the organizers when they made the request, but he has done his best to comply by suggesting a few topics of interest.

Any scientific meeting that takes a limited region of the earth's crust as its purview ought to begin its proceedings by considering whether that region has valid limits. At the outset, we can agree that the atmosphere itself is unlikely to provide us with any boundaries for the polar regions. Obviously we are concerned in this symposium with the poleward flanks of the westerly belts, and with the complex, irregular cores of the circumpolar vortices. But no meteorologist will be able to say with any conviction that fixed limits for the polar regions can be identified within the westerly belts. The whole tendency in modern discussions of the general circulation is to consider at least whole hemispheres, which are thought of as indivisible. Atmospheric boundaries—fronts, in other words—are occasionally visible, but they are ceaselessly in motion. If one prepares a frontal and airmass climatology, one can perhaps identify, as Bergeron did, average positions for the polar and Arctic fronts, which define Arctic and Sub-Arctic domains within the atmosphere, with equivalents in the Southern Hemisphere. But again the modern tendency is to discount the significance of these frontal boundaries, and to emphasize instead the exchanges of mass and momentum between the polar regions and the westerlies. We can, therefore, expect little help from the synopticians and the dynamic meteorologists in our search for meaningful limits on the polar maps.

Moreover, to underline this conclusion, we can remind ourselves that the working methods of mid-latitude analysis can be carried to the poles in a

manner quite impossible in the equatorward direction. Tropical meteorologists have ready-made boundaries for their areas of specialization; the failure of the geostrophic approximation as the equator is approached necessitates very different methods of analysis, and the more or less fixed position of the sub-tropical highs effectively separates the tropical specialists' sphere of interest from mid-latitudes. Nothing of the kind is true near the poles. The Coriolis parameter, and hence geostrophic control, is a maximum at the geographical poles, and ordinary synoptic analysis works right across the Arctic. Even the analytical models of fronts, airmasses and frontal cyclones can be identified at the North Pole, though there is plenty of evidence that ordinary frontal analysis has been pushed to excess in high latitudes, as Reed suggests later in this symposium. But the north polar region, at least, forms a continuum with the adjacent westerlies, and there is no profit in trying to find any artificial atmospheric frontier between them. The same may not be true of the Antarctic, where the unique physiography of the continent plays a vital role. Even here, however, it is the exchanges across the Antarctic limit that matter to the meteorologist, rather than that limit's position.

Lastly, in this recital of negatives, we shall make the point that the geographical poles themselves are far from being satisfactory centers for the polar belts, in the meteorological sense, since the polar climates are by no means symmetrical about the earth's axis, especially in the Northern Hemisphere. Hence latitude is not a very meaningful basis for the delimitation of Arctic and Sub-Arctic. Certain physical controls are, of course, latitudinally distributed. Thus the Coriolis parameter, and its first space derivative, are obviously simple functions of latitude. So also is the annual cycle of solar radiation. On the other hand, the earth's surface is by no means symmetrical to the polar axis, nor is the rhythm of outgoing terrestrial radiation. Hence both the circulation of the atmosphere about the polar regions and the distribution of the physical climatic elements show eccentricity to the poles, marked in the Northern Hemisphere, less so in the south, where the center of gravity of the Antarctic landmass is not far from the South Pole itself.

Nothing displays this eccentricity more vividly than the mean 500 mb chart for the Arctic winter. It is at once apparent that the circumpolar westerlies are at present distributed about a twin-centered vortex, with centers over northern Siberia and the Canadian Arctic, the latter being actually very close to the geomagnetic pole, a coincidence noted with interest by Flohn and Scherhag in recent discussions of the northern climates.[3]

It is thus apparent that the dynamic and synoptic aspects of meteorology are going to provide no simple means of delimiting the polar regions. In fact the view has been expressed to the author that this symposium is merely concerned with an area for which synoptic data have hitherto been lacking, i.e. that the Arctic and Antarctic, meteorologically speaking, are only areas of ignorance, like much of the South Pacific. This view he cannot share, in spite of the difficulties raised above.

Climatological Definition of the Arctic and Sub-Arctic

The author's purpose in this paper is to suggest that we can define valid limits for the Arctic and Sub-Arctic, as well as for the subdivisions of the

latter, on the basis of certain quasi-stationary geographical distributions that are in partial equilibrium with the present climate. Among these distributions we can select the following as being of greatest value:

(i) The zonation of natural vegetation, with corresponding soil conditions, energy and moisture balances, all of which are eccentric to the geographical pole.

(ii) The occurrence overland of permanently frozen ground ("permafrost") and associated "patterned ground" phenomena.

(iii) The distribution of sea, lake and river ice throughout the year.

(iv) The remaining traces of the once-extensive continental glaciers.

(v) The circulation of Arctic and Sub-Arctic water masses.

Obviously an adequate discussion of all five of these involves interdisciplinary gymnastics of a kind few meteorologists will relish. Moreover, some of them have been exhaustively reviewed by world authorities in the recent literature —permafrost by MULLER,[6] glaciers by FLINT[2] and AHLMANN,[1] and the circulation of northern water masses by the chairman of this symposium, Dr. H. U. Sverdrup. The author has himself[4] discussed point (i), the circumpolar distribution of vegetation, in a paper accessible to most meteorologists (but familiar to few).

What is the significance of these distributions to the meteorologist interested in the north? Probably the answer is three-fold. Firstly, all five of them drastically affect the nature of the earth's surface, and all five, therefore, have a bearing on airmass modification, in the synoptician's terms, and on the earth's energy and moisture transformations. Secondly, each of them is in some sort of equilibrium with the contemporary dynamic climate of the Arctic and Sub-Arctic. Thirdly, all of them have shifted geographically in the recent past in response to shifts in the atmospheric circulation and thermal régime. The study of climatic change, apart from the recent climatic fluctuation, still depends very largely on our ability to recognize the earlier circumpolar distributions of glacial ice, vegetation, cold ocean water and patterned ground. Clearly, then, the five distributions listed above are of real interest to the meteorologist, even if his physical training equips him poorly to study and understand them.

It is noteworthy that these arguments make sense only in the Northern Hemisphere. In the south, open ocean supervenes where in the north comes the forested Sub-Arctic. In the sense of the present paper, there is no "Sub-Antarctic"; the ice-covered Antarctic continent is in direct and roughly latitudinal contact with the Southern Ocean. What the author has to say is in the main relevant only to the Arctic, which is fortunately of greater interest than the Antarctic to the organization sponsoring this symposium.

In effect, then, the author proposes that the Arctic and Sub-Arctic can be defined only in climatological terms, using this to mean the understanding of the mechanism of climatic control over the five distributions listed. We are a long way from understanding completely how this control is exerted. The glaciologists, arguing from accumulated observation on the ablation and replenishment of existing glaciers, have made a good beginning as far as land ice is concerned. Qualitatively, at least, there is also some knowledge of the relations between existing climates and the circulation of Arctic and

Sub-Arctic seas. But as far as vegetation is concerned—and this, in the author's view, is the most significant distribution to the meteorologist—we have hardly got beyond the stage of noticing geographical coincidences between temperature and vegetation distributions, of the kind popularized by Köppen, and recently revived by the present author.[4] Vegetation is often described as the great climatic integrator. Few meteorologists probably realize how little actual evidence and how much inelegant theory lies behind that statement.

What is needed, in this last field, is accurate, meteorologically-inspired research into the actual interchanges of moisture and energy between the atmosphere and the different northern vegetation types—research on Geiger's microclimatological or Thornthwaite's topoclimatological scales. The great belts of vegetation that encircle (at least on land) the northern pole differ radically in albedo, moisture requirements and surface roughness. Anyone who has worked in the north knows that there are wide changes in surface wind régime (and hence blowing snow), low-level turbulence and even in characteristic low cloud forms, as one goes from the Arctic seas on to the tundra land, and from tundra into Sub-Arctic forest. These differences, of considerable synoptic importance, are obviously due to variations in the albedo, temperature, wetness and roughness of the earth's surface. To determine how these properties vary will help the northern forecaster, and at the same time will provide the data for a better understanding of the climate-vegetation relationships discussed above.

Research along these lines has been begun by the author and his colleagues at the McGill Sub-Arctic Research Laboratory, at Knob Lake, in the Labrador Peninsula. The vegetation of the entire peninsula has been classified physiognomically, and its distributions mapped. The characteristic albedo and annual moisture régimes over the different types are now being investigated instrumentally. Later we may attempt wind profile measurements over the same surfaces. We hope that this work can ultimately be made to bear fruit both ecologically and synoptically.

Present-day Extent of Arctic and Sub-Arctic

Finally, in order that the published proceedings of the symposium may contain a map with surface detail, the author presents *Fig. 1*, on which most of the five distributions listed above are combined and extended.

In sea areas, *Fig. 1* shows the normal extent of ice-affected waters in mid-winter (February–March) and late summer (August). In winter the greater part of the indicated area is frozen sufficiently solidly to make the physical characteristics of the sea surface similar to—though not identical with—a snow-covered land surface. Breaks exist locally, as in the famous "North Water" of Baffin Bay, but their effects on the atmosphere are fairly local. In summer, however, the vast areas of sea surface still containing thawing ice (including the so-called "permanent-pack" of the Arctic Ocean) constitute giant thermostats tied to the freezing point. The map, drawn with modifications from the U.S. Hydrographic Office's Ice Atlas of the Northern Hemisphere,[8] is still rather approximate.

Over land, the distributions shown in *Fig. 1* are of two sorts. The treeline, north of which lie the treeless tundras, high in albedo and low in roughness,

is a limit that has been mapped with varying degrees of detail all round the pole. In the forested Sub-Arctic, or *Boreal* zone, however, world maps of structure still await completion. Instead the map shows the approximate extent of three thermally defined zones, shown by the author[5] to be closely correlated with forest physiognomy in areas where reasonable maps exist. The defining parameter is C. W. Thornthwaite's potential evapotranspiration,[7] an accumulating function of air temperature. Zone I, from the treeline

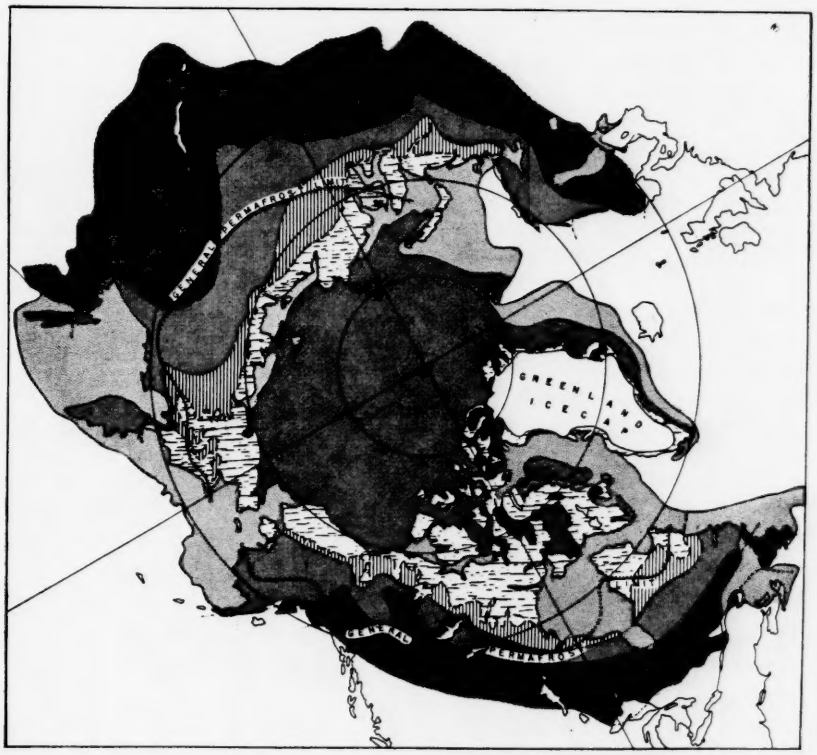


Fig. 1. Some geographical distributions under climatic control

to a value of 35 cm, is associated in the mapped areas with intermingled tundra and thin woodland, mostly confined to the lowest ground. Albedo remains fairly high, but roughness is only a little higher than in the tundra. Zone II corresponds to an open structured woodland in most areas, with an abundant lichen floor having an albedo almost as high as dirty snow.

Windspeeds are much lower than in Zone I. Zone II extends between 35 and 42 cm, on Thornthwaite's parameter. Zone III, corresponding to a closed-crown forest, usually conifer-dominated in the areas mapped, has a much lower albedo than the other zones, and windspeeds are very low except above the forest crown.

The distribution of general permafrost is also indicated in *Fig. 1*, again rather tentatively; patchy permafrost occurs over wide areas south of this limit. The general permafrost limit is quite strikingly dissimilar to the thermal zones of the Boreal zone in its course around the pole. It appears to follow fairly closely the mean annual air isotherm of 23°F.

The map also shows the surviving continental glaciers.

Fig. 1 is, of course, merely a record of present-day geographical distributions, insofar as we know them. The idea of discovering their climatic controls, put forward here as the real basis of definition for Arctic and Sub-Arctic, has proved so intractable and so elusive to the research worker that we can still, in 1956, barely hint at what these controls may be. Accordingly, the author has to admit that he finishes this paper with a confession of past failure, not only by himself, but by most physical geographers and biologists. To rectify this state of affairs, it is imperative that the physical meteorologist get to work on the problems presented by *Fig. 1*. Obviously these problems can be solved only by physical methods in conjunction with the specialist skills of the botanical ecologist, the physical geographer and the glaciologist. And the great gap in Arctic physical research has been the almost total lack of meteorological interest in these questions. One may number on the fingers of one hand the names of the meteorologists who have contributed significantly in this field. The great pioneer work of Sverdrup on the polar pack-ice, and the contributions of Wallén, Liljequist and Orvig on the equilibrium of existing glaciers, are rare examples of what is needed in the other fields listed above.

References

[1] Alhmann, H. W-son "Glaciological Research on the North Atlantic Coast" *R. Geogr. Soc. Research Series: No. 1*, 1948

[2] Flint, R. F. *Glacial Geology and the Pleistocene Epoch*, pp. 32–64; John Wiley & Sons, New York, 1947

[3] Flohn, H. "Zur Aerologie des Polargebiete" *Met. Rund.* Vol. 5, 1952, pp. 81–87, 121–128. (Available in English translation from Arctic Meteorology Research Group, McGill University, Montreal, Canada.) Flohn refers to Scherhag's views

[4] Hare, F. K. "Some Climatological Problems of the Arctic and Sub-Arctic" *Compendium of Meteorology*, Amer. Met. Soc., pp. 952–964, 1951

[5] ——"The Boreal Conifer Zone" *Geographical Studies* Vol. 1, pp. 4–18, 1954

[6] Muller, S. W. *Permafrost;* Ann Arbor, Edwards Bros., 1947

[7] Thornthwaite, C. W. and Hare, F. K. "Climatic Classification in Forestry" *Unasylva* Vol. 9, pp. 50–59, 1955

[8] U.S. Hydrographic Office. *Ice Atlas of the Northern Hemisphere;* Washington, U.S. Navy. Charts 1–12, 1946

Suppl.
Part. I
1956

SIMILARITIES AND CONTRASTS BETWEEN THE ARCTIC AND ANTARCTIC MARINE CLIMATES

RICHARD M. HOLCOMBE*
U.S. Navy Hydrographic Office, Suitland, Maryland

With the increase in the number of scientific and exploratory expeditions to polar regions during this century, new information has given scientists an opportunity for re-examination of many of the effects of the air circulation both at the surface and aloft. Continental and maritime influences are partially responsible for many climatic differences between the Antarctic and Arctic. In the last few years the U.S. Navy Hydrographic Office has been able to delineate some of these northern and southern hemispheric comparisons with information gathered from many countries. With the increase in available punch-card data many presentations such as frequency distributions of various weather elements have been made possible during the past 15 years.

By assuming a model consisting of a polar stereographic map projection of the Arctic and a mirror image of a polar stereographic map projection of the South Pole for the Antarctic, and by using similar scales for both projections, a visual comparison of climatological conditions for both hemispheres is facilitated. This model has the convenience of counterclockwise circulation about lows, and clockwise circulation about highs in both hemispheres. Also, statistical comparisons of basic meteorological parameters including their deviations are readily obtained from this model.

The Hydrographic Office also has managed to obtain summaries of raobs and rawins in the Arctic and Antarctic regions. The combined results of HIJUMP and the Byrd Expeditions indicate close agreement in respect to general circulation with the findings of Dr. Arnold Court and some of the ideas postulated after the Scott Expedition. The extremely broad ranges of height and temperatures for standard levels above the tropopause at Alert, NWT, do present some questions.

I. INTRODUCTION

SINCE this is an international meeting, it should be emphasized that this paper would have been impossible without the co-operation of many nations. By combining more than 10 marine punch-card decks, comprising the efforts of more than 5 countries, and supplementing this data with the information obtained during polar expeditions, the U.S. Navy Hydrographic Office has been able to amass, with the co-operation of the National Weather Records Center at Asheville N.C., a comprehensive world-wide coverage of marine meteorological information. In 1953 there were over 288 million punch-cards stored at Asheville. Of this number, about 25·5 million are marine observations.

Until very recently, most of our knowledge about the polar regions has been deduced from information obtained from the periphery latitudes. This paper is no exception. A complete description of the polar climatic conditions and the environmental influences on man is beyond the scope of this short paper.

* The opinions expressed in this paper are those of the author as an individual and are not to be construed as necessarily reflecting the official views of the Navy Department.

However, it is hoped that the type of material shown will indicate the quantity and variety of material that is presently available for research purposes.

The basic model for comparing graphically the Arctic and Antarctic marine climates is a pair of superimposed polar map projections of similar scale. All Arctic information is presented in blue or black and all Antarctic information is presented in red or orange. A dashed line, regardless of color, represents a wintertime phenomenon (February or August, depending on the hemisphere), whereas a solid line represents a summer-time condition. In order that the east and west directions can be shown as the same in both hemispheres, the Arctic is viewed from above the earth's surface and the Antarctic is viewed from the center of the earth. Another advantage of this graphical method of presentation is that it permits cyclonic circulation to be shown as counterclockwise in both hemispheres.

II. Comparison of the Physical Features

Fig. 1 shows that Antarctica's 5·5–6·0 million square miles of land comprise about the same area as the Arctic seas. In brief, Antarctica is said to be a landmass surrounded by water, whereas the Arctic is said to be a water mass surrounded by land. However, there are indications that the average person tends to underestimate the continental extent of Antarctica. Antarctica's 6000 ft mean altitude coupled with 1·9 million square miles of pack-ice area, even in the relatively warm month of February, compose one of the largest and most pronounced continental régimes in the world. However, the degree of continentality of the ice area in the Southern Hemisphere has not been ascertained as yet. In summer, the area south of the pack-ice limits of Antarctica (7·3 million square miles) is slightly larger than the total area of South America. In winter, when there is an additional 3·2 million square miles of pack-ice, the total area south of the pack-ice limits (10·5 million square miles) is about 1·5 million square miles larger than the North American landmass, including Mexico, the United States, Canada, Alaska and Greenland. The increase in pack-ice area (approximately 3·2 million square miles) between summer and winter is about two tenths of a million square miles more than the total area of the continent of Australia.

The first sign of persistence throughout the year of the Antarctic climatic régime is seen in an average winter-summer ice-limit difference of only 6·1° of latitude compared to 12·2° for the Northern Hemisphere.

The Arctic Ocean's area of approximately 5,400,000 statute square miles, being about the same size as the Antarctic Continent, is about the only similarity in respect to topographic characteristics between the two polar regions. First, the average elevation in the Antarctic is about 6000 ft above mean sea level compared to approximately sea level for the Arctic. Second, there is a significant source of heat energy below the ice surface in the Arctic. It is hoped that the figures shown will reflect the influence of the topography on the climate in terms of data now available.

III. Surface Meteorological Comparisons

A. Temperature

Fig. 2 is intended to show winter and summer extremes for the location of the 32°F isoline in both the Northern and Southern Hemispheres. The

Suppl.
Part.I
1956

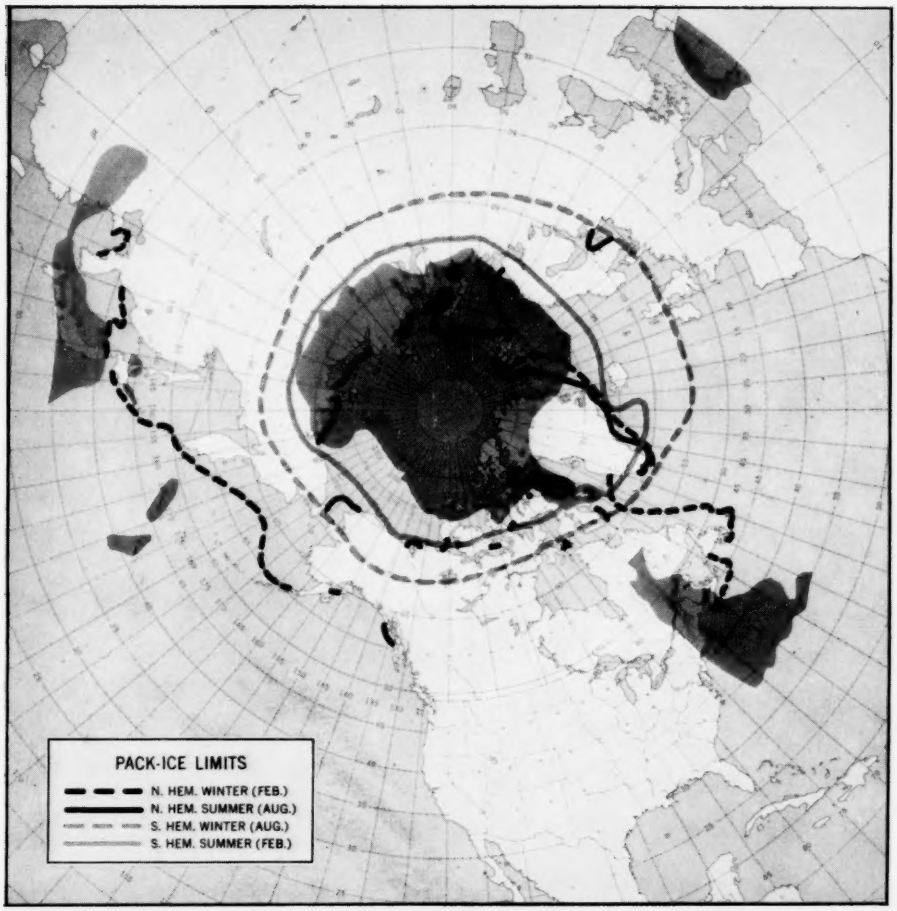


Fig. 1

(*Facing page* 10)

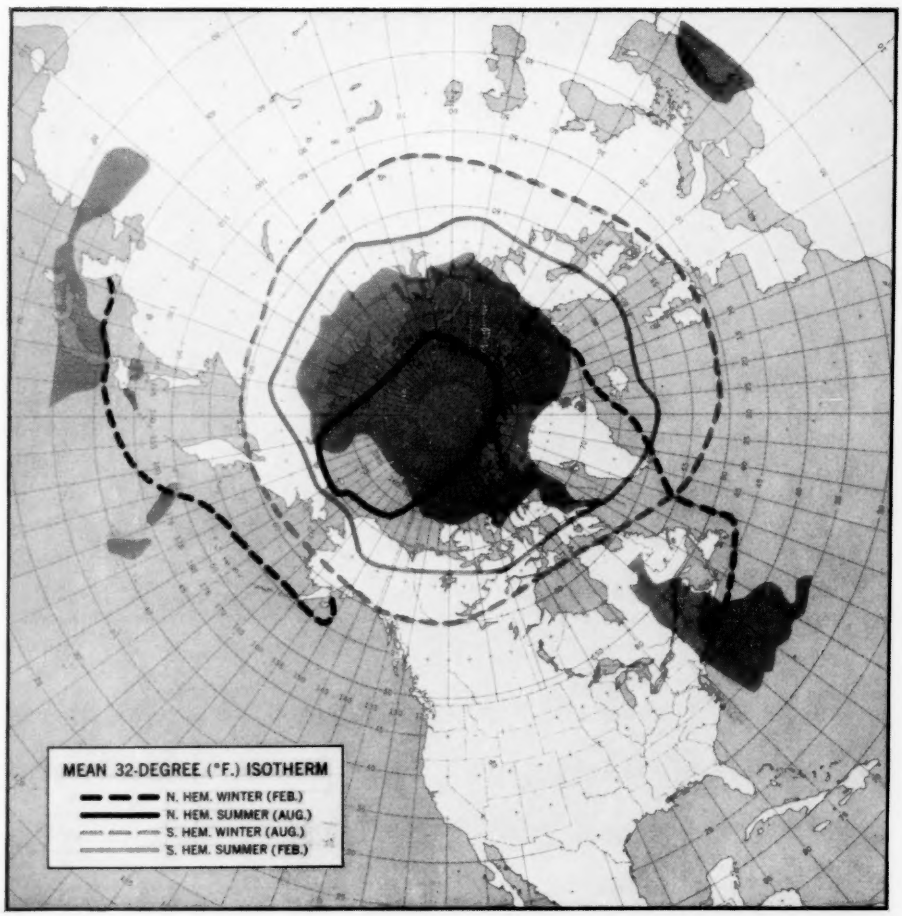

MEAN 32-DEGREE (°F.) ISOTHERM
N. HEM. WINTER (FEB.)
N. HEM. SUMMER (AUG.)
S. HEM. WINTER (AUG.)
S. HEM. SUMMER (FEB.)

Fig. 2

Suppl.
Part.I
1956

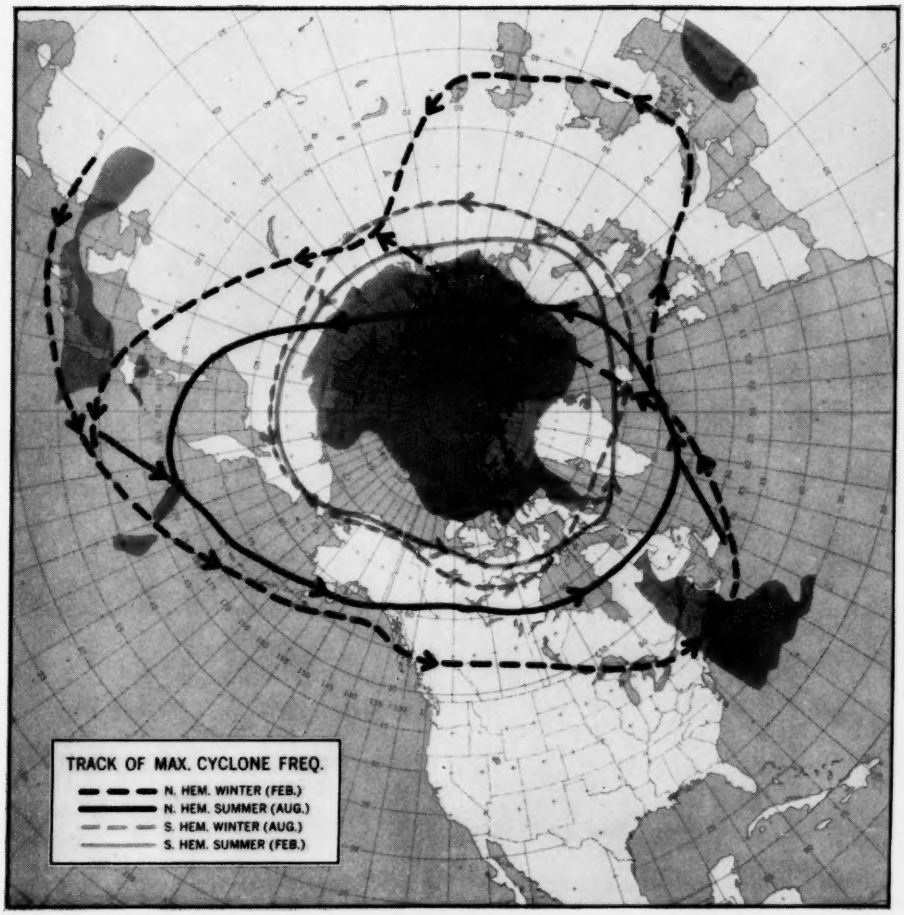


Fig. 3

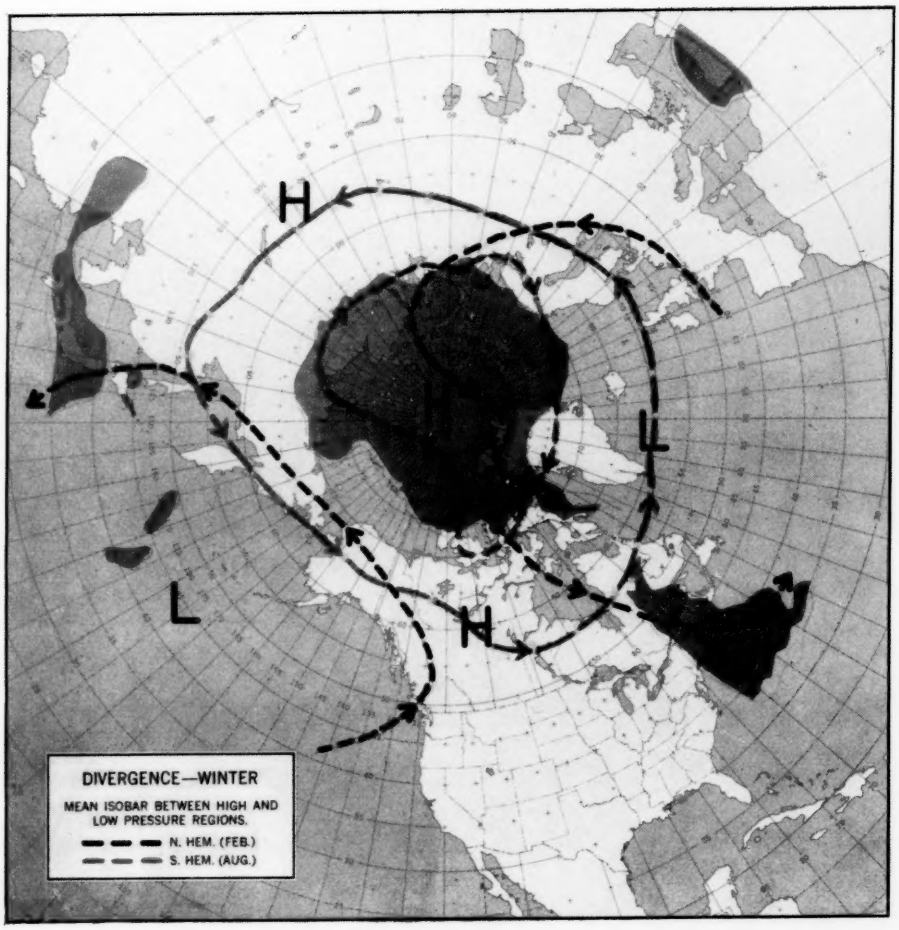


Fig. 4

Supp
Part
1956

Suppl.
Part. I
1956

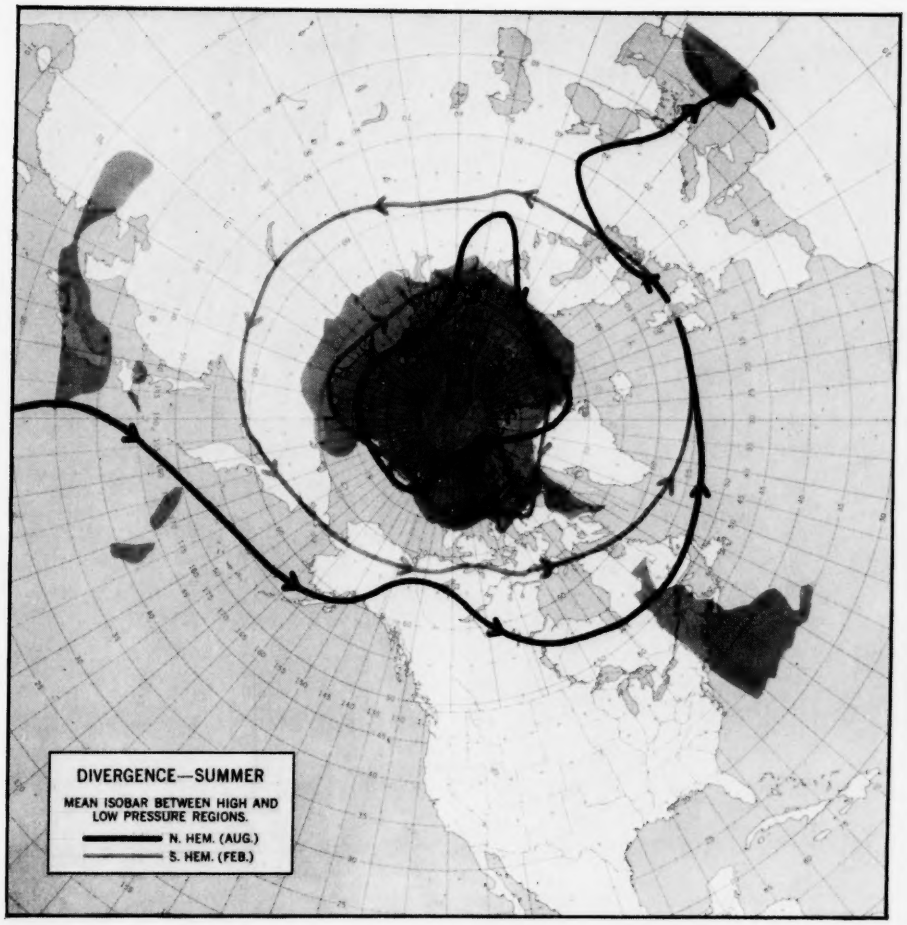


Fig. 5

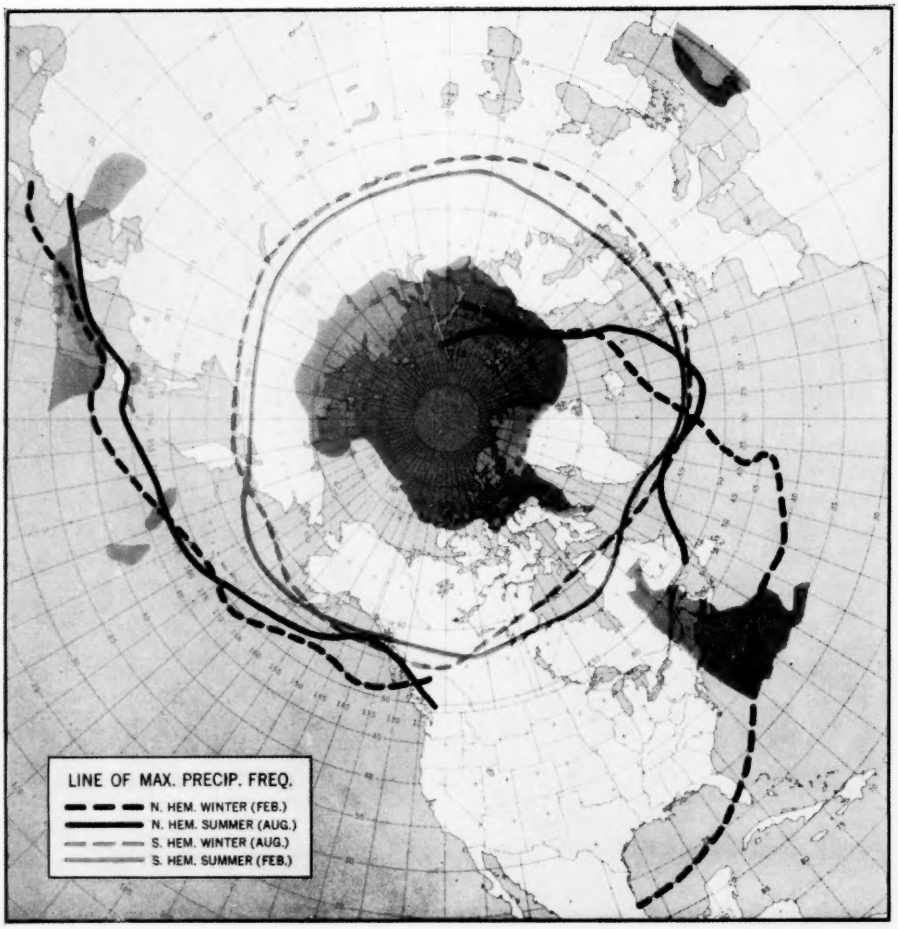


Fig. 6

(*Facing page* 11)

dashed and solid lines indicate winter and summer phenomena in their respective hemispheres. The concentric symmetry in the Antarctic is persistent throughout the year. The average latitudinal difference between winter and summer is only 7 degrees. In summer the mean latitude is 64°S, and in winter the mean latitude is 57°S. In the Northern Hemisphere, the average difference in latitude is 22 degrees between a mean summer latitude of 78°N and a mean winter latitude of 56°N. The annual range of the mean monthly 32°F isoline varies between 51° and 61°S in the Southern Hemisphere compared to 40° and 84°N in the Northern Hemisphere. As a result, the Antarctic experiences the coldest summers in relation to the rest of the world, and there are insufficient data at present to judge whether the Antarctic is colder than the Arctic during the winter. The lack of symmetry in the Arctic is indicated by −90° and −87°F being recorded in Siberia and Greenland, respectively. The world's unofficial lowest temperature of −108°F, recorded in Siberia, is 25°F lower than the minimum recorded in the Antarctic.* However, the estimated lowest annual average temperature of −30°F is in Antarctic's Adelie Land. The off-pole centering of the summertime 32°F isoline in the Arctic reflects the maritime influence of the North Atlantic Ocean and the North Atlantic Drift, and also the relatively southward displacement of the northern coasts of Siberia and Alaska.

Suppl.
Part. I
1956

B. Cyclones

Fig. 3 again reflects the year-long persistence of conditions in the Antarctic compared to those in the Arctic. In the Southern Hemisphere, the winter-summer difference in the mean latitude of the track of maximum cyclone frequency is 1·9 degrees in the Southern Hemisphere and 5·4 degrees in the Northern Hemisphere. The mean latitude of this zone of maximum cyclone frequency in the Southern Hemisphere is 66° and 64°S for summer and winter, respectively, compared to the Northern Hemisphere's 62° and 57°N for summer and winter, respectively. Since it was intended that this paper be confined to the environments of polar regions and surrounding areas, the track of relatively high cyclone frequency in winter along the northern Mediterranean has been discounted. The frequency of systems passing along this southern track is about equal to that along the track north of Europe. Also, the branch off the coast of Japan in both summer and winter has been discounted.

In order that the relative frequencies of cyclones between the Northern and Southern Hemispheres may be more readily compared, it should be mentioned that the average winter and summer frequency of cyclones passing the 0°, 90°E, 180° and 90°W meridians is about the same in both hemispheres. In winter, the average monthly (February or August, whichever applies) frequency of cyclones crossing each of these meridians is 13 for the Northern Hemisphere and 14 for the Southern Hemisphere. In summer, the average monthly frequency of cyclones crossing each of these meridians is 9 for the Northern Hemisphere and 10 for the Southern Hemisphere. The latitudinal scatter of cyclones is indicated in the following manner: in the Southern Hemisphere, two-thirds of the 5-degree quadrangles reporting cyclones along

* On 17 September, 1957, at the South Pole's Amundsen-Scott Station, a temperature of −102·1°F was recorded.

a meridian have a monthly frequency of about 2 compared to slightly more than 1 for the Northern Hemisphere. The mean monthly range of frequency along the track of maximum frequency varies between 5 and 2 for the Southern Hemisphere and 1 and 4 for the Northern Hemisphere. The relatively high amount of scatter of cyclone tracks in the Northern Hemisphere could be an indication of the relative degree of forecasting difficulty between the two hemispheres.

C. Regions of divergence

Figs. 4 and *5* delineate the regions of divergence and convergence in the two hemispheres. As in the previous figures, the Northern Hemisphere is indicated by blue or black, and the Southern Hemisphere is indicated by either red or orange. The first of these figures, with the dashed isolines, applies to winter only. These isolines represent the mean isobar between the high and low pressure regions in their respective hemispheres. In the Southern Hemisphere, this value was estimated to be 995 mb compared to 1015 mb for the Northern Hemisphere. The estimation of the mean sea-level pressures in Antarctica required considerable interpretation. Since the map for the Southern Hemisphere is the mirror image of a polar stereographic projection for the Arctic, the circulation about the low in both hemispheres is counterclockwise.

In winter the continents of Eurasia, North America and Antarctica are all dominated at the surface by a region of divergence. Only in the Antarctic is this region of polar divergence centered in the proximity of the pole. Surrounding Antarctica is an elongated belt of convergence. Surrounding the Arctic, the region of convergence is confined primarily to the marine areas.

In summer, the polar divergent areas are relatively similar. This seasonal trend has many similar characteristics to that of the mean monthly 32°F isotherms discussed above. However, even in summer the Arctic region of divergence is centered off the pole partly as a result of the monsoonal effect of Eurasia, partly as a result of the Arctic Ocean being off center from the North Pole, and partly as a result of the influence of the North Atlantic Drift.

D. Precipitation

Fig. 6 is intended to show by graphical means a comparison in maritime precipitation conditions in the higher latitudes of the two hemispheres. The legend, in keeping with the general format of the previous figures, indicates Northern Hemispheric conditions in blue or black, Southern Hemispheric conditions in red or orange, winter conditions by dashed lines, and summer conditions by solid lines. The isolines are intended to indicate the latitude of maximum precipitation frequency in representative months during the extreme seasons of the year (February and August). As in the previous figures, this one brings out the general year-long persistence in the Southern Hemisphere. The mean winter and summer latitudes of the maximum precipitation are 56° and 57°S, respectively, for the Southern Hemisphere, and 45° and 47°N, respectively, for the Northern Hemisphere. However, in the Northern Hemisphere the continental influence in the western Atlantic, the effect of the North Atlantic Drift in the eastern Atlantic, and the continental influence in the western Pacific induce marked seasonal contrasts.

In terms of the percentage of observations reporting precipitation of all types, the mean frequencies along the zone of maximum frequency for winter and summer is relatively high in the Southern Hemisphere with 54 and 43 per cent, respectively, compared to 32 and 23 per cent for the Northern Hemisphere. However, the interior of Antarctica is considered to be nearly arid. The standard deviation of precipitation frequency along the zone of maximum frequency in the Southern Hemisphere is only 2 and 3 per cent for winter and summer, respectively, compared to 11 and 6 per cent in the Northern Hemisphere for winter and summer.

E. Gale frequency

Showing significant wind conditions in terms of gale-force winds is not very representative of actual conditions because of the local influence of topography, and the nearshore land-sea effect. Along the coast of Antarctica, gale-force winds have caught ships off guard during clear weather. Five to 10 miles off shore or up the coast, the wind would be calm. Gravity winds of this sort almost cost the lives of a shore party during the past summer expedition to Antarctica.

Suppl.
Part.I
1956

Since most of the available marine observations of gale-force winds are made in regions where the normal pressure gradient is relatively steep, the data is biased. For this reason no figures were prepared. Typical examples of localities where the frequency of gale-force winds is high are the screeching fifties and sixties in the Southern Hemisphere, and the regions of semi-permanent lows in the Northern Hemisphere.

In the Southern Hemisphere the mean latitude of the zone of maximum frequency of gale-force winds varies between 54°S in winter and 56°S in summer. This range is relatively small compared to 62°N and 46°N for the corresponding seasons in the Northern Hemisphere. Mean monthly extreme latitudes of the zone of maximum frequency of gale-force winds are 47°S and 60°S in the Southern Hemisphere and 16°N and 76°N in the Northern Hemisphere. The frequency of gale-force winds along the zone of maximum frequency is 24 and 17 per cent for winter and summer, respectively, in the Southern Hemisphere, and 24 and 5 per cent in the Northern Hemisphere.

IV. Upper Air

A. Little America

Figs. 7 and *8* are designed to show a visual comparison of the wind and temperature conditions aloft at Little America. The dashed line is the 270°A dry adiabatic lapse rate. The dashed-dot line represents the temperature-height lapse rate of the Standard Atmosphere. The solid line represents the mean temperature-height lapse rate for the month of February. The numbers to the right of the temperature-height lapse rate curve represent the mean monthly dew-point temperature for the various standard levels; and the numbers to the left of the curve represent the mean monthly departure of the heights (feet) of the standard levels from the heights as indicated by the Standard Atmosphere. This graph would indicate that during the summer there is a tendency for the base of the tropopause to be located around the 300 mb level.

The top set of the pairs of numbers along the right edge of the graph indicate, for each standard level, the resultant wind direction and speed in knots and the bottom set the scalar windspeed in knots. The direction is the resultant direction from which the air has moved. In the relatively stable surface layers, the southeasterly resultant direction conforms with the properties of the cold surface high in which the resultant circulation is away

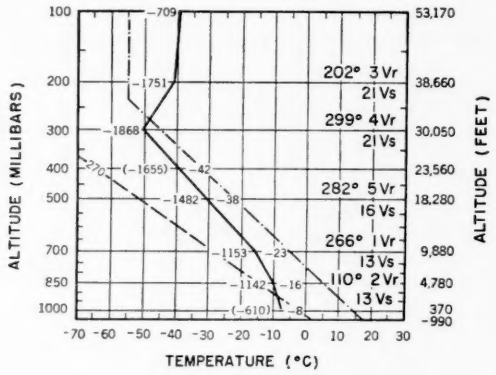


Fig. 7

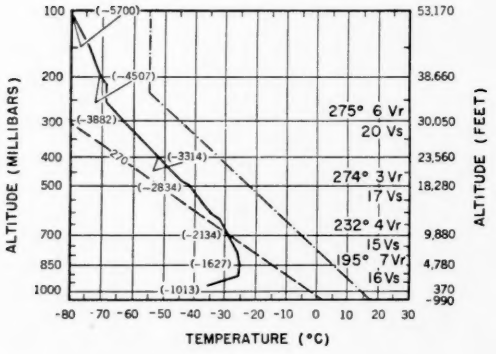


Fig. 8

from the pole. Between the 700 and 300 mb levels, the resultant flow toward the pole reflects the presence of a low aloft. This circulation characteristic was first realized during the Scott Expedition, when it was noted that the wind on the top of the erupting volcano Mt. Erebus was blowing in a direction opposite to the wind at the surface. The relatively low ratio between the resultant and scalar winds in the vicinity of the 700 mb level reflects the presence of a transition zone between the easterlies associated with the surface high and the westerlies aloft. The sampling size of the south-southwest wind at the 200 mb level is too low to be of significance.

In *Fig. 8*, the legend is essentially the same as in *Fig. 7*. The marked surface inversion is typical of wintertime conditions. Surface winds for Little

America, not shown on this graph, are for the most part from the east. The resultant flow from the pole extends to a level closer to the 500 mb height than indicated by the summertime data. However, the most significant feature is the absence of a tropopause. This winter characteristic was first treated in detail by Dr. Arnold Court from the data compiled during the Byrd Expeditions.

B. Alert, NWT

Figs. 9 and *10* are designed for the purpose of showing simultaneously the frequency distributions of pressure altitudes and temperatures for specified standard levels and the general relationships of these meteorological elements

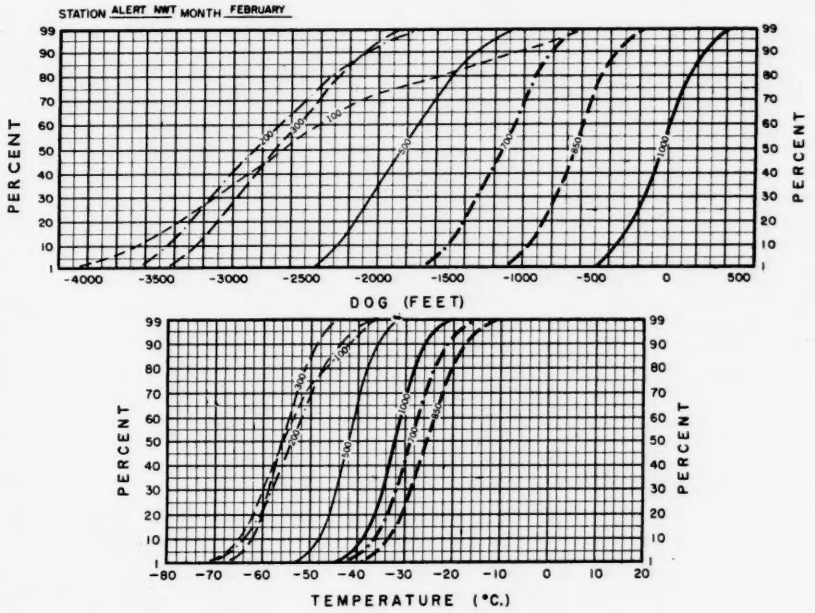


Fig. 9

between one standard level and another. The heavy lines are for the standard levels 700 mb and below. The thinnest dashed line represents the 100 mb level. At Alert in February, there is a strong tendency for an inversion level between 850 and 700 mb. The base of a normally well-defined tropopause is in the vicinity of the 300 mb level. The appearance of a tropopause during a season when there is a general absence of one in the Antarctic is explained by the greater extent of latitudinal exchange of air in the Arctic as compared to the Antarctic. The Dog scale represents the difference in altitude in feet between the reported values for standard levels and the heights of the standard levels of the standard atmosphere. The most significant feature in this figure is the wide range of reported altitudes for the 100 mb level. At the 100 mb level, the range is 3400 ft compared to an average range of 1250 ft for the remaining standard levels.

Suppl.
Part.I
1956

In summer, the average range in heights of standard levels is about 1200 ft for all levels including the 100 mb level. However, the intense surface inversion does not prevail. The bunching of the pressure-height distribution curves is an indication of a tendency toward a relatively stable

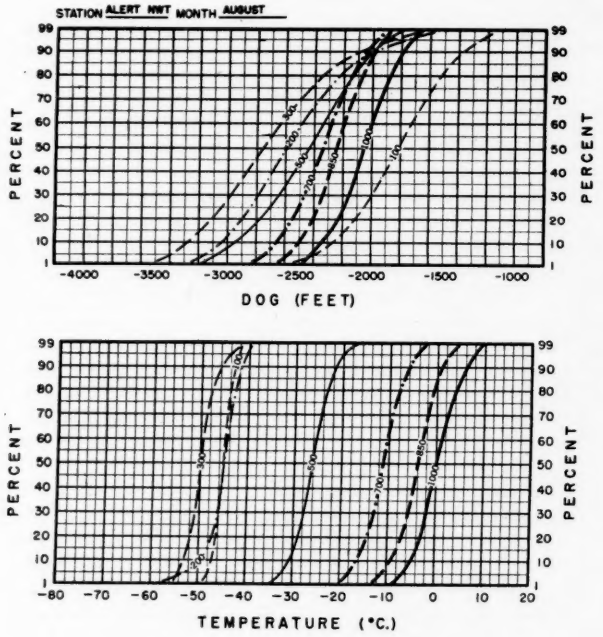


Fig. 10

atmosphere above the 700 mb level in comparison to winter conditions. The base of a relatively well-defined tropopause is in the vicinity of the 300 mb level.

V. Conclusion

Until recently, many of the conclusions concerning conditions in the vicinity of the poles have been estimated from our knowledge of the behavior of the weather in the regions surrounding the poles. In the Southern Hemisphere, meteorologists have had to rely on the ship reports of whaling vessels and expeditions in the vicinity of 60°S. The coastal weather data has been either of relatively short records, or confined to localized, widely distributed sites, i.e. Heard Island and the Palmer Peninsula. As for the Arctic, most of the data is confined to widely scattered stations along the Arctic Ocean. Data compiled from occasional expeditions to the Pole, Arctic flights and ice islands are extremely valuable. However, normals, departure from normal and persistence are difficult factors to determine. It is hoped that "on the spot" observations taken during the coming Geophysical Year will supply the answers to many questions.

References

BROOKS, C. E. P. and CARRUTHERS, N. *Handbook of Statistical Methods in Meteorology;* Met. Office, London, M.O. 538, 1953

COURT, ARNOLD. "Antarctic Atmosphere Circulation" *Compendium of Meteorology*, pp. 917–941; Boston: Amer. Met. Soc. 1951

——"Tropopause Disappearance during the Antarctic Winter." *Bull. Amer. met. Soc.* Vol. 23, No. 5, pp. 220–238, 1942.

MASSACHUSETTS INSTITUTE OF TECHOLOGY. "Southern Hemisphere Weather Maps, July 1948 to June 1950." On microfilm at the Hydrographic Office

MAURY, M. F. *The Physical Geography of the Sea and its Meteorology*, 8th ed.; Harper and Brothers, New York, 1868

RUBIN, M. J. "Seasonal Variations of the Antarctic Tropopause." *J. Met.* Vol. 10, No. 2, pp. 127–134, 1953

SVERDRUP, H. U. Meteorology, Part I: Discussion, *The Norwegian North Polar Expedition with the "Maud", 1918–1925, Scientific Results* Vol. 2, pp. 1–331, 1933

UNION OF SOUTH AFRICA, WEATHER BUREAU, DEPARTMENT OF TRANSPORT. *Synoptic charts for the Southern Hemisphere, 1951–1953.* Pretoria, Film, Unpublished

U.S. HYDROGRAPHIC OFFICE. Climatological Data for Various Marsden Squares. Tabulated by the U.S. Weather Bureau Decks: 115 (U.S. Merchant Marine Deck, 1941–1948); 116 (U.S. Merchant Marine Deck, 1949–1953); 192 (Hamburg, D. Seewarte Deck); 193 (Nederlands, Met. Instituut Deck); 194 (Great Britain, Met. Office Deck); and 195 (U.S. Navy Ships' Logs Observations Deck). Unpublished

——"Sailing Directions for Antarctica Including the Off-lying Islands South of Latitude 60°" *Hydrographic Office Publication* No. 138; Washington, 1943.

U.S. NAVY. CHIEF OF NAVAL OPERATIONS. FLIGHT SERVICE SECTION. AEROLOGY. "Aerological Observations and Summaries for the Antarctic, from 15 December 1946 to 15 March 1947; a project of Operation Highjump, Task Force 68" *NAVAER* 50–1R–214; Washington, 1948

WOODWARD, R. S. "Smithsonian Geographical tables" *Smithsn. misc. Coll.* 854, 1929

Suppl.
Part.I
1956

THE RECENT VARIATIONS OF THE CLIMATE AT THE NORWEGIAN ARCTIC STATIONS

TH. HESSELBERG and T. WERNER JOHANNESSEN
Det Norske Meteorologiske Institutt, Oslo, Norway

The Norwegian Meteorological Service has erected five meteorological stations in the Atlantic sector of the Arctic regions; namely, Spitsbergen (1912), Jan Mayen (1920), Bjørnøya (1921), Myggbukta (1922) and Hopen (1946).

The observations from these stations have been used for a study of the recent climatic changes. Of special interest are the data from Spitsbergen where the series of observations go back to 1912. During the first years the observations show no conspicuous climatic change, but then comes a rapid rise of the temperature in the years 1917 to 1922. The increase of the mean temperatures in this period was about 7 degrees Celsius in the winter, 3 degrees in the spring, 2 degrees in the summer, 3 degrees in the autumn and 4 degrees for the whole year. After the year 1922 the temperature continued to rise until the war broke off the series, but the rise was much slower. The temperature increased in these years about 1 degree Celsius. After the war the temperatures are a little lower, the temperature rise thus seems to have culminated.

Before the establishment of the meteorological station in 1912 we have only sporadic observations of the temperature and they are not fit for a numerical study of the variations of the climate.

There are, however, rather many observations of the glaciers and it can be ascertained that there has been a considerable melting of the glaciers so that several of them have retired considerably, and this retreat began before 1912. The present amelioration of the climate in the Arctic area therefore must have begun before the year 1912. There are evidences that seem to indicate that it began in the 1870-years, approximately at the same time as the temperature rise began in more southerly regions.

The rise of the temperature in Spitsbergen is large compared with the rise in other parts of the world (about five times as great as in Norway). This fact can be explained by the position of Spitsbergen at the southern border of the inner Arctic area. This inner zone is retreating toward north in connection with the general heating, and Spitsbergen that formerly was situated north of the path of the cyclones surrounding the inner Arctic zones is now more frequently visited by cyclones that bring mild air from south and south-west. Of special importance is that the augmenting cyclonic winds sweep away the surface layers with very cold air that in calm weather is produced by the outgoing radiation.

For the study of the climatic changes it is of special interest to follow the variations in the Arctic where they are so great and also to follow the contractions and extensions of the Arctic zone. It is therefore a great advantage that in the later years a series of meteorological stations has been erected around the Arctic Ocean, and for the Geophysical Year 1957–58 further improvements of the network of observations have been planned.

1. THE STATIONS

FOR the benefit of the forecasting service in Norway, and especially for the storm-warnings along the coasts, the Norwegian Meteorological Institute has erected several weather stations in the adjacent Arctic regions, namely

Supp
Part
1956

Spitsbergen (established in 1912), Jan Mayen (1920), Bjørnøya (1921), Myggbukta (1922) and Hopen (1946).

The position of these stations are given in the map (*Fig. 1*). Meteorological observations are made three to eight times per day at the synoptic hours and weather telegrams are regularly sent to Oslo, from where they are issued

Fig. 1. Position of stations in the Arctic

for international use. The observations have further been used for calculation of climatological data which since 1913 have regularly been published in the Norwegian meteorological yearbooks.

On the map in *Fig. 1* is also seen the position of the other synoptic stations north of 70°N, that sent reports in January 1956. Almost all of them are rather new stations and therefore still of small value for the study of climatic changes.

2. The Climatic Conditions

Before we discuss the variations of the climatic conditions, a short description of these conditions might be of interest. In the table some climatic data

Some Climatological Data from the Years 1946–54

Station	*Winter*					*Summer*				
	Average temperature	*Average temperature min*	*Average frequency of days with*			*Average temperature*	*Average temperature max.*	*Average frequency of days with*		
			Clear sky	*Fog*	*Wind force ≥ 9B*			*Clear sky*	*Fog*	*Wind force ≥ 9B*
Myggbukta	–20·7°	–41·8°	23·9	3·6	4·9	2·3°	18·1°	13·8	35·2	0·4
Isfjord R.	–9·8	–26·4	9·2	0·3	1·8	3·5	11·0	4·2	20·3	0·0
Hopen	–10·1	–29·0	9·4	4·7	5·3	1·4	9·5	3·0	52·3	0·2
Jan Mayen	–3·6	–17·0	1·0	8·8	14·2	4·3	13·4	0·8	38·8	1·1
Bjørnøya	–5·6	–21·3	2·6	2·9	6·9	3·6	14·8	2·2	43·8	0·1
Vardø	–4·0	–16·0	1·9	5·2	3·1	8·0	20·5	3·6	18·0	0·0

from the after-war period 1946–54 are given for our five Arctic stations. For comparison the corresponding data are also given for the town Vardø, which is well known for its rough weather.

In the first part of the table we find data from the winter months December to February, and in the second part the corresponding data for the summer months June–August. The first column gives the average temperatures, and the second column the average minimum and maximum temperatures, respectively. The next columns show the mean frequency of days with clear sky, fog and stormy weather (wind force 9 Beaufort or more).

From the table we will see that the winter is very cold in Myggbukta where the climate has a rather continental character, but it is also cold in Isfjord R. and in Hopen. There are many days with clear weather in Isfjord R. and in Myggbukta, and few days with fog and stormy weather. The winters are characterized by a clear sky and cold, relatively calm weather. Almost all the days with fog are concentrated in the summer, and clear days are then seldom.

Quite different are the weather conditions at Jan Mayen and Bjørnøya, both situated in the low-pressure region surrounding the interior Arctic regions. They are frequently visited by cyclones. The temperature is therefore relatively high in winter, while the summer is cool. There are few clear days here and many days with stormy weather and fog. The storms are especially frequent at Jan Mayen, where wind velocities of 50 m sec^{-1} or more have been measured many times. At Bjørnøya there is less stormy weather, but more fog.

Upon the whole the weather conditions are not agreeable in the stormy belt surrounding the interior Arctic region.

Suppl.
Part. I
1956

3. The Homogeneity of the Observation Material

With special regard to the climatological studies the stations have of course been maintained as far as possible at the same places and with unaltered conditions for the observations. But some changes have been necessary.

Thus the Spitsbergen station had to be removed in 1931 from Green Harbour to Longyear City, where it remained until the new Isfjord Radio had been established at Kapp Linné in 1935, about 15 km from the previous station in Green Harbour. These changes at the Spitsbergen station have caused inhomogeneity for all meterological elements except for the atmospheric pressure and the temperature. For these elements it was possible to combine the different series into one, valid for the actual station Isfjord Radio.

During the war all our Arctic stations were destroyed, Jan Mayen and Myggbukta in 1940, Bjørnøya and Isfjord R. in 1941, and all of them except Jan Mayen were out of function until the end of the war.

The station at Jan Mayen was re-established a few months after the destruction, but not at the former place on the east coast, but in a small valley in the middle of the island. After the war the station was moved several times. It is now situated about 6 km from the old place. For the atmospheric pressure and the temperature these small displacements were insignificant, but for all the other meteorological elements the homogeneity was destroyed.

The other Arctic stations could not be re-established until the end of the war: Myggbukta and Isfjord R. at exactly the same places as before, while the station at Bjørnøya had to be removed 7·5 km from the old place. For all these three stations the observations of temperature and pressure after the war can be considered as a continuation of the series before the destruction of the stations. There remains, however, a gap of 3–4 years in the series.

4. The Series of Observation

The International Meteorological Organization has (Warsaw 1935) defined the climate as the mean meteorological conditions calculated over a period of 30 years, and the variations of the climate as the variations of these means.

From our stations in the Arctic region we have no unbroken series of observations for a period of 30 years. The longest one is from Isfjord R. where we have observations from 1912 to 1940, a gap in the years 1941–45 and observations again from 1946 onwards.

For all the other stations the series of observations are considerably shorter, and we are therefore obliged to base our studies of the mean conditions on periods considerably shorter than 30 years. In the following we will use periods of 1, 5 and 10 years.

It is a disagreeable fact that we can only use the observation series of atmospheric pressure and of air temperature for our studies because there have been changes at the stations that have made the series of the other meteorological elements (precipitation, humidity, wind, etc.) inhomogeneous. It is, however, no serious obstacle to our studies, because we know, to some extent, how the other elements vary when we know the variation of the most important elements: the atmospheric pressure and the air temperature.

Finally, it must be remarked that we have no possibilities for the computation of the mean values for the period 1901–30 which generally is used as reference period. As the observations at Spitsbergen began in 1912 we have, however, been able to compute mean values for the period 1912–30, and these years were then chosen as reference period for our stations in the Arctic region.

It is of course a disadvantage not to be able to use the common reference period, but in reality it does not matter much what period we choose. In fact only the absolute values of the deviations from the mean values are modified by the choice of reference period while the variations themselves are independent of this choice.

In spite of the deficiencies of the series of observations the material can be used, and it gives a clear picture of the variations of the temperature conditions in the Atlantic sector of the Arctic region that is covered by the Norwegian meteorological stations.

5. The Variations of the Temperature Conditions

Experience derived from studies of the secular variations of the temperature in other parts of the world has shown that it is of no special value to examine the general features separately for each month. It is even an advantage to augment the smoothing by using the means for 3 months instead of those for each month. We will therefore limit ourselves to the study of the mean

temperature in the different seasons (December–February, March–May, June–August and September–November) and for the year.

The computed tables show that the mean temperatures have increased at all stations and that the rise is greatest in the winter and very modest in the summer. Of special interest are the data from the Spitsbergen station, which was erected in 1912, some years before the great change in the temperature conditions took place. This happened in the years 1917–22.

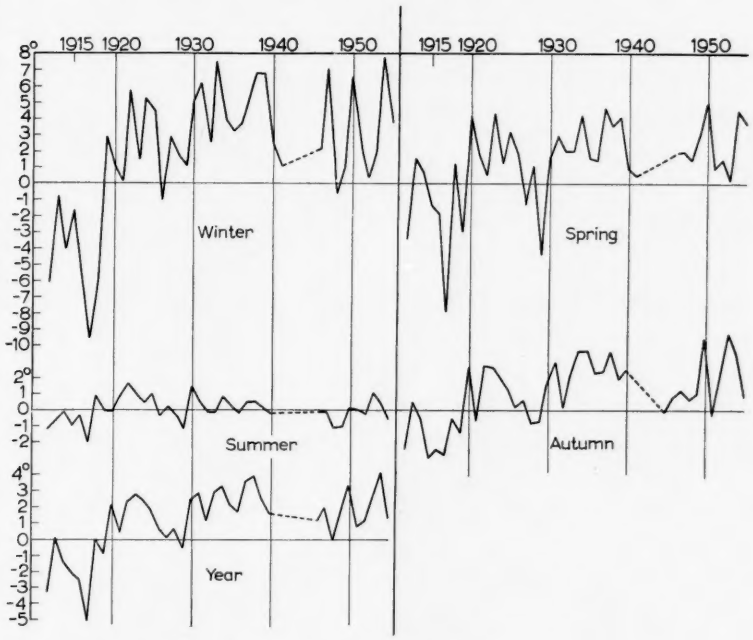


Fig. 2. Mean temperature at Isfjord R.

The remarkable rise in temperature is clearly seen in *Fig. 2*, that gives the departures of the mean temperatures from those of the reference period at Isfjord R.

A smoothing by hand gives a rise of about 7 degrees for the winter and later on a slow increase of about 1 degree. For the other seasons the sudden rise around the year 1920 is approximately 3 degrees for the spring, 2 degrees for the summer, 3 degrees for the autumn and 4 degrees for the whole year.

Figs. 3 and *4* show the average temperature departure for the seasons and for the year, smoothed over 5 years. The curves give the consecutive 5 year arithmetic mean departures from the mean values in the reference period 1912–30.

We see that the rise in the temperature at Isfjord R. is confirmed by the data from Vardø. Also here the increase is especially rapid about the year 1920, but the rise is more modest. For the other stations the observations begin later, when the rapid temperature increase has been replaced by a

Suppl.
Part. I
1956

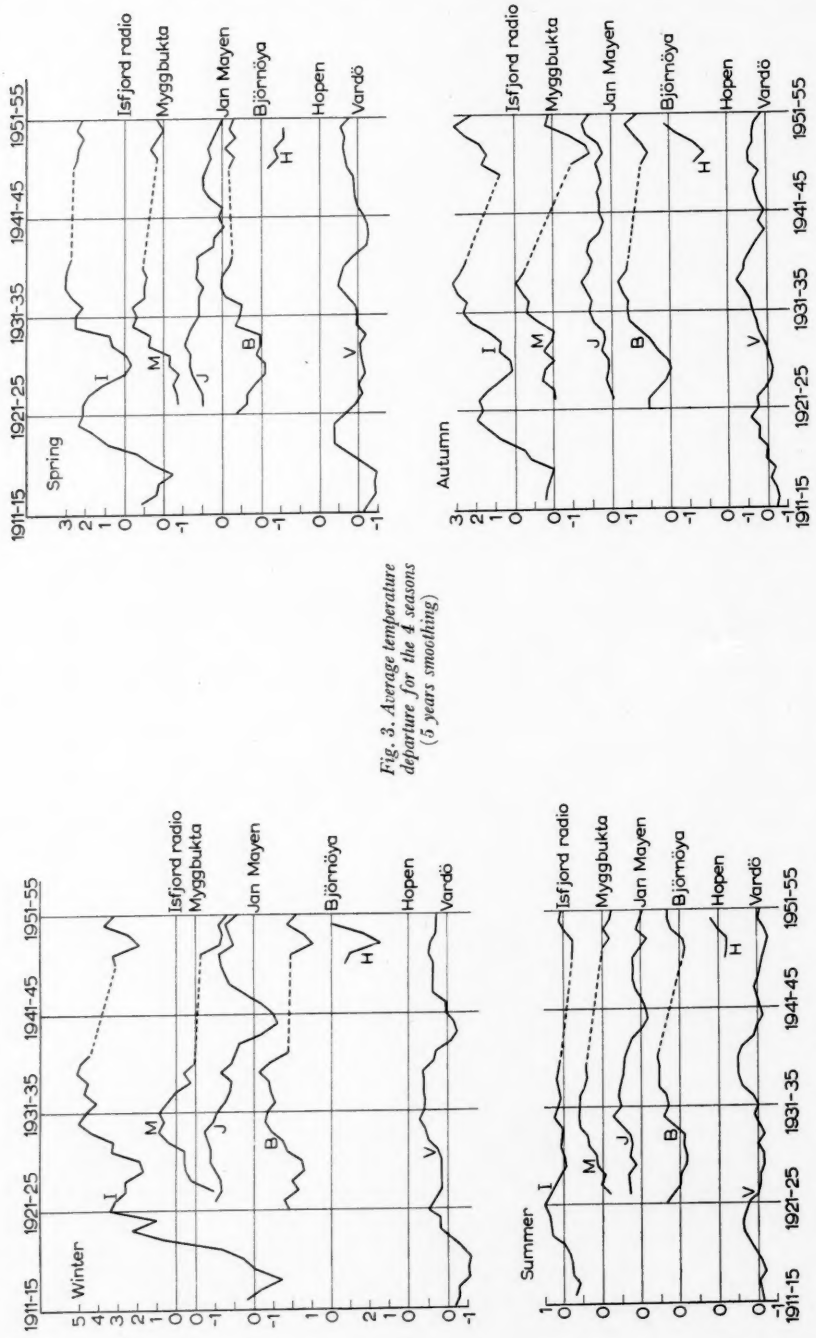


Fig. 3. Average temperature departure for the 4 seasons (5 years smoothing)

slow increase. We find this moderate increase common for all our stations, except Jan Mayen, until the 6 years cessation during the war. In our curves for the 5 year means this stop causes a break of 6 + 5 years. After this stop the temperatures are a little lower than before the war. The temperature rise in the Arctic thus seems to have culminated.

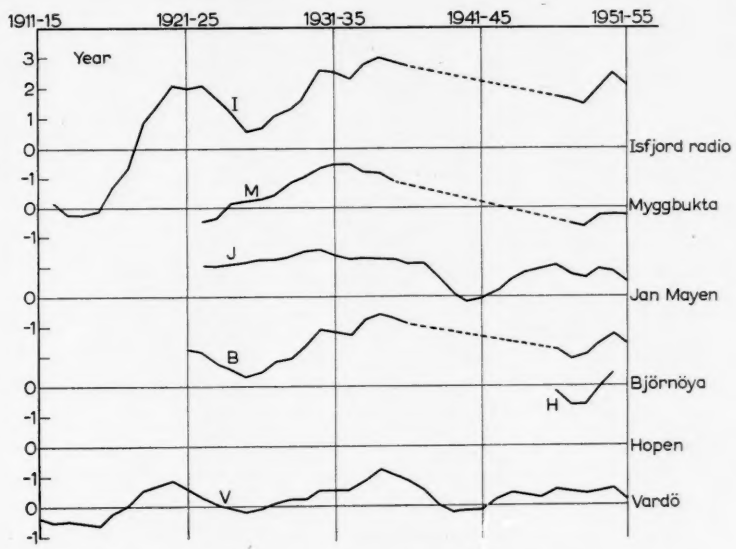


Fig. 4. Average temperature departure for the year (5 years smoothing)

In *Fig. 5* we find the average temperature departures at Isfjord R. after a smoothing over 10 years. The rapid change about 1920 is because of the smoothing artificially spread out over a greater interval of years, and the break because of the war gives a gap of 6 + 10 years. In order to avoid these disadvantages we based our discussions on the departures with 5 years smoothing.

6. Some Consequences of the Improved Climate

With the rising air temperatures the sea-ice in the Arctic region has decreased and the ice-front has retired to the north.

A practical consequence of this fact is a prolongation of the season for shipping between Norway and Spitsbergen. How closely the length of this season is correlated with the temperature is seen from *Fig. 6* where the upper curve indicates the length of the shipping season and the curve below the mean temperature for the year, both smoothed over 5 years.

Another practical consequence is the change at the banks west of Bjørnøya and Spitsbergen. Up to the year 1917 investigations showed that there was little fish, but from the year 1925 commercial fishing has been carried on with good results; fishing which has extended almost over the whole year.

The most conspicuous change at Spitsbergen is the great melting of glaciers. In *Fig. 7 (a)* we see a picture from Raudfjord (79° 40′N) taken in

Suppl. Part. I 1956

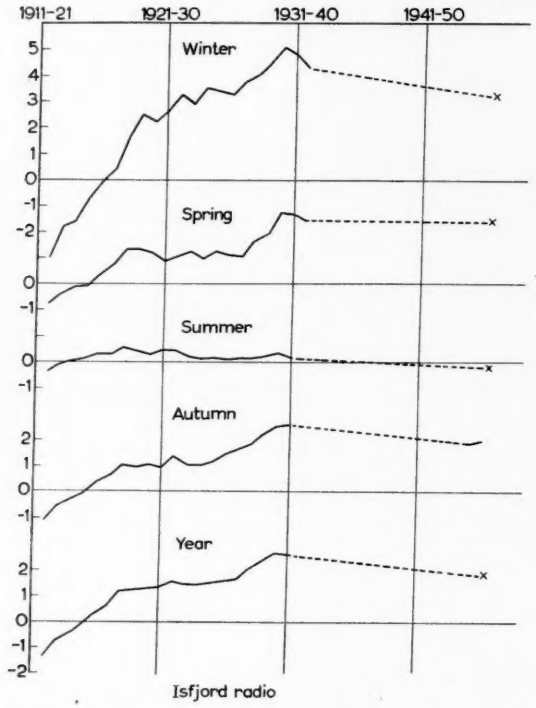


Fig. 5. Average temperature departure for the 4 seasons and the year at Isfjord R. (10 years smoothing)

August 1909. In *Fig.* 7(*b*) we have a photograph taken from the same place 27 years later (August 1936). The retirement of the glacier front in the fjord is of 1250 m, and the mountains show how the glaciers have diminished.

In *Fig. 8* we see a graphical representation of the retirement of the Paula Glacier (77° 50′N). Of special interest is the fact that the retreat of the

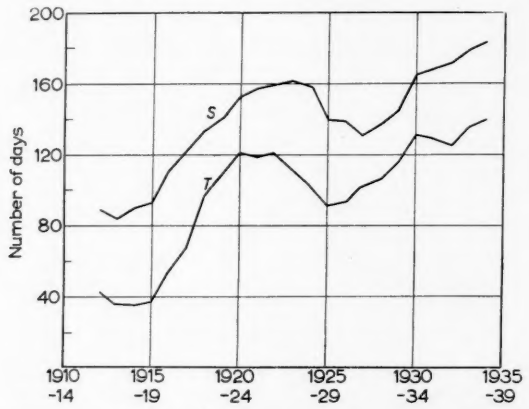


Fig. 6. Relation between the length of the shipping season, and the mean temperature (5 years smoothing)

Fig. 7 (a).

Variation of glacier front at Raudfjord, Spitsbergen, from August 1909 *(a) to August* 1936 *(b)*

Suppl.
Part. I
1956

Fig. 7 (b)

(*Facing page* 26)

Supp
Part
1956

glacier began before 1913. This fact and similar evidences from other glaciers show that the improvement of the climate in Spitsbergen did begin before 1912, the year of establishment of the regular meteorological observations. It seems probable that it began about the year 1870, together with the recent

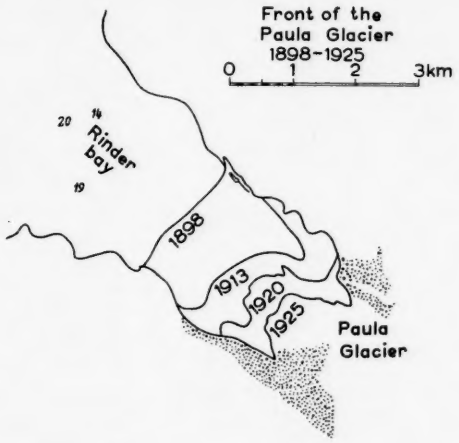


Fig. 8. Variation of front of the Paula Glacier, Spitsbergen. Rinder Bay to the left

temperature rise in Northern Europe. A basic for this supposition is that the number of days with Arctic sea-ice at the coast of Iceland began to decrease considerably at that time.

7. Discussion

In 1943 Hesselberg and Birkeland published a paper on the secular variation in atmospheric pressure and wind. From this paper we reproduce a map (*Fig. 9*) that shows the increase of the mean atmospheric pressure for the winter from the period 1912–21 to the period 1926–35.

We see that the pressure has increased 4·5 mb over the British Isles, while it has fallen with 2·5 mb at Spitsbergen. Over the Arctic region the isallobars run from south-west to north-east, and this means that we have got an additional wind from south-west. These additional winds bring more heat to the region, and are one direct cause of the rising winter temperatures.

It may further be mentioned that the isallobars show that the belt of low pressure surrounding the interior Arctic region has moved to the north and that the cyclones have got more northerly paths. At Spitsbergen there are therefore more winds than before and the number of calms have diminished. It decreased from 33 to 21 per cent from the period 1914–18 to the period 1924–28. In the cold seasons these augmenting winds sweep away the surface layers with very cold air and give considerably increasing temperatures.

The sudden rise of the temperature about the year 1920 can thus be explained by the fact that the cold surface layers to a great extent are swept away by the augmenting winds.

Seen in relation to the general circulation the isallobars show that the Arctic High has been pressed to the north in the Atlantic sector, so that the inner Arctic region has diminished.

The corresponding isallobaric maps for the other seasons show augmented southerly winds in the spring and in the autumn, while the additional winds are weak and easterly in the summer. This is in accordance with our temperature curves that show a considerable increase of temperature for both the spring and the autumn and a small rise of the summer temperatures.

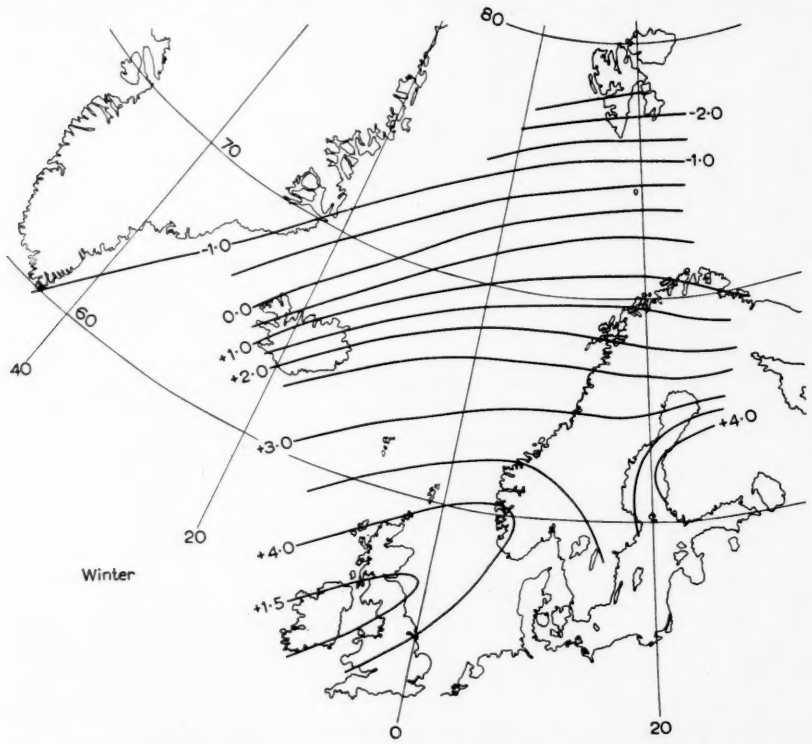


Fig. 9. Increase of mean atmospheric pressure in the winter from the period 1912–21 to the period 1926–35

In regard to the summer temperature it is further to be remarked that a transport of warm air from the south or even an increase of absorbed heat cannot bring the mean summer temperature at the surface essentially above zero, because the heat to a great extent will go to melting of the ice.

The above considerations show that the changes in the temperature conditions in the Arctic region can be explained from the variations in the pressure field and the wind field. These changes are consistent with each other and the same is certainly the case with the variations of the other meteorological elements. But these conformities in the variations do not give the cause of these variations.

We might give the question a general form by asking why the general circulation has been intensified with rising temperatures, such as we have

found over Northern Europe and the part of the Arctic region that is covered by Norwegian stations.

In this connection it is natural to mention the hypothesis of Nansen and Helland-Hansen. Already in the year 1916 they published the idea that variations in the radiation of the sun should be the initial cause for the climatological variations on the earth. We know, they said, that the sun is not an invariable star and that the radiation varies so that the energy absorbed by the atmosphere also must be variable. In periods with augmenting energy received, the atmosphere as a whole gets higher temperatures. But the atmosphere is an engine. With more energy received the engine will go faster or, in other words, the general circulation must increase in order to smooth out the increased difference of heating at higher and lower latitudes.

Over some parts of the earth there will be more wind from lower latitudes —as the case has been over the North Atlantic and the Atlantic part of the Arctic region—so that they get a relatively large rise of the temperature. Over other parts of the earth there may be an increase of the wind from higher latitudes and here the flow of colder air can be so great that the cooling becomes greater than the heating.

At winter solstice the difference of heating at low and high latitudes has its maximum effect and the consequence of an increase of heat received from the sun should therefore also have its maximum at that time. This is what we have found, the temperature rise has its greatest value for the winter and the autumn.

Suppl.
Part.I
1956

Upon the whole the ideas of Nansen and Helland-Hansen suit so well to the meteorological data that it seems justified to consider them a good working hypothesis.

Important studies of the changes of the climate have in the later years been made by different meteorologists and most of these investigations have been based upon ideas similar to those of Nansen and Helland-Hansen. It would go too far to discuss the different papers; it should, however, be mentioned that Simpson in 1934, from his studies of the ice ages, came to the conclusion that the meteorological effect of a variation in the radiation from the sun should be especially great in the Atlantic sector of the Arctic area. We should here expect considerable contractions and extensions of the inner Arctic region and corresponding displacements to the north and to the south of the tracks of the cyclones; that is in good agreement with the later observational data. Arctic observations are thus of special importance for the study of the climatic variations.

In the map in *Fig. 1* we saw the Arctic Ocean in later years has been surrounded by a series of weather stations and further improvements of the network of stations have been planned for the International Geophysical Year, especially for the upper air.

On the other side the interest of astronomers for the study of the variations of the radiation from the sun has got an impulse from these climatological studies and it seems reasonable to hope that the continued efforts will give results that can give a full explanation of the climatic variations and open the way for long-range forecasts.

A BRIEF REVIEW OF RESEARCH IN ARCTIC METEOROLOGY

A. D. BELMONT
McGill University, Montreal*

THE purpose of this review is to point out briefly some of the results which have been found in the past, to mention what is being done in the United States and Canada, to evaluate the research data available, and to suggest two major problems which could be investigated with existing data. We will think of the Arctic for our purpose to be mainly the Arctic Ocean, Greenland and the adjoining coastal belts. Although much can be learned about the Arctic by comparison with the Antarctic, we will restrict ourselves mainly to the north polar regions and to problems of special importance there.

The past history of Arctic meteorological research may be divided into that before and that since the last war, corresponding to the advent at that time of regular aerological data from the Arctic. The first phase succeeded in establishing many facts concerning the surface climatology and, to a limited extent, the structure of the lowest layers of the atmosphere. The present aerological period is attempting to describe the general circulation at upper levels over the Arctic Basin, and to understand its causes.

This problem was also one of the reasons behind the First Polar Year of 1882–83, which succeeded in extending the observational limits farther toward the north, and made possible, so far as I can discover, the preparation of the first circumpolar maps (Vincent). There followed, in the next 50 years, many expeditions to the Arctic, of which only a few had scientific results.

We can list the First and Second Norwegian North Polar Expeditions with the *Fram*, 1893–96 and 1898–1902, whose meteorological results were prepared by Mohn, and which gave us the first direct observations over the Basin. The famous Norwegian North Polar Expedition with the *Maud*, 1918–25, although not so fortunate as the *Fram* in its drift area, contributed, in the meteorological results prepared by Prof. Sverdrup, some of the best observed and analyzed data for the Arctic Ocean. The greatest contribution was the first aerological information (from kites) of the inversion structure over the pack-ice. It seems appropriate this year that we should also mention that exactly 50 years ago this summer, Hergesell, on an expedition sponsored by the Prince of Monaco, took balloon observations over the Norwegian Sea as far north as Spitsbergen. These supplied some of the first evidence of the maritime inversion. The Second Polar Year, 1932–33, again enlarged the surface station network to enable some idea of the climate for previously little-known sectors. Also, its data were compiled to prepare the set of historical Northern Hemisphere maps published by the Deutsche Seewarte. Several meteorological expeditions to Greenland made upper-air observations, of which the most successful were the German one at Eismitte 1930–31,

* Now with Atmospheric Section, General Mills, Minneapolis, U.S.A.

and the French operation at the same location 1949–52. Naturally, only the latter expedition was able to take radiosonde ascents. From its first presentation in 1910 until today, the "glacial anticyclone" theory has been a subject of controversy. The results of the Greenland expeditions can be used to disprove the permanent "glacial anticyclone" there, by evidence of cyclonic activity across central Greenland. In 1937–38 the Russians took the initiative in a direct approach to the general circulation of the polar area by landing a scientific group by airplane near the North Pole and letting them drift on an ice floe for nine months into the Norwegian Sea. In the excellent report of the results, Dzerdzeevskii conclusively proved that their sector had indeed been affected by typical cyclonic weather and that no permanent polar anticyclone existed.

During and after the last war the observing network in the north increased rapidly. Ten years of radiosonde data now exist for Thule, the first of the jointly operated weather stations in the North American Arctic. Five other joint stations now operate in the Canadian Archipelago, and Denmark has greatly increased the radiosonde network around Greenland. In 1947 the United States Air Force started the "Ptarmigan" weather reconnaissance flights from Fairbanks to the North Pole at the 500 mb level, and in 1950 three dropsondes were released regularly along the flight, giving us the first upper air data over the central Arctic up to 500 mb. The year 1952 was another landmark in Arctic data coverage. The U.S.A.F. landed on T-3, one of several ice islands in the Basin, and for two years regular radiosonde observations plus surface records were taken. The station was evacuated when it drifted too near Alert to warrant the difficulties of maintaining it. From then to the present there has been an ice island or ice floe weather station in operation somewhere over the Arctic Basin, for the Russians have also established semi-permanent stations on ice floes, whose duration lasts until the floe breaks up or drifts too far south.

The next chance for improving the Arctic network will be the International Geophysical Year, although it does not appear that much is planned with respect to Arctic upper-air measurement. The year will be more important for the Antarctic than the Arctic.

II

I would like to discuss more fully the research which has been done and is being done in North America with the Arctic data now becoming available. Until 1949 results of expeditions were prepared by individual scientists and there was no organized meteorological research devoted primarily to the Arctic area. About 1949 it was decided to initiate a long-term group project to attempt to find out more about Arctic circulation with the available data. There were at that time no satisfactory polar area synoptic charts which could permit any confidence in the apparent development of pressure patterns. Movements of systems can be so rapid that once-daily charts may give an entirely erroneous impression of the synoptic situation development during the intervening 24 hr. Operational maps drawn under short-time limits, for limited sectors, and which frequently miss data due to communication difficulties, were too incomplete. The U.S. Northern Hemisphere Historical Map Series as prepared at that time was many years behind, and

Suppl.
Part. I
1956

was drawn for a period when good Arctic data did not yet exist. It was therefore decided to collect original records for 1948—the most recent year for which good data were available then—and prepare a detailed series of surface and upper-air charts using the thickness method and greatest possible care in all steps. This was the first goal of an Arctic meteorology project established at the University of California at Los Angeles at the end of 1950. Although most data were stored at the U.S. Data Collection Center, it was not possible to get the coded reports in the same form as on the teletype, suitable for plotting, so that work had to be done from enlarged microfilm of original station records. Further, Russian radiosondes had to be individually recomputed to find the 850 mb level which was not reported. We prepared four surface maps at the standard hours, two intermediate surface maps to agree with the upper-air map hours, and 1000, 850, 700 and 500 mb charts. By the time the first month, July 1948, was completed, it was realized that the process was too slow and the series was discontinued.

At the same time, an analysis of the available radiosonde data from BW-1 at the southern tip of Greenland, for 1948, was carried out. This station's data were unique at that time since it took observations at hours which corresponded to local noon and local midnight and, in addition, took two intermediate observations a day. The results of this study showed a most surprising conclusion: that the mean monthly temperatures for constant pressure surfaces appear to be warmer at midnight than at noon, in the stratosphere, in late winter and spring. We have not been able to explain why this should be, nor have we been able to get more years of data from the same station to check these results. Published Thule data for 1946–49 showed the same result although the local hours were 1000 and 2000. Radiosonde radiation errors were not corrected for, but these ascents took place in total darkness at Thule. Furthermore, correcting for the radiation error would increase the observed effect.

Since almost nothing was written in English about Arctic synoptics, we found it worth while to translate several papers and a book by Dzerdzeevskii giving the results of the Russian North Pole Expedition of 1937–38. As they have been published during and just after the war, they were largely unknown in America.

The next paper to emerge from the UCLA project was a study of the dropsondes, for the first two years, 1950–52, of the "Ptarmigan" flights. The soundings were grouped into 4 latitudinal belts from 70° to 90°N and analyzed with respect to the surface inversion, following Sverdrup's classic example. "Ptarmigan" soundings were all more smooth in appearance than those of the *Maud* Expedition, but a discussion of the results will be included in one of the papers to be given later in this symposium and so will be omitted now.

The last report of the UCLA series was to compare the first winter of ice island T-3 stratospheric temperatures with other polar stations. It appeared that T-3 had an upper atmosphere more similar to Alaska than to the nearby Canadian or Greenland stations. This may have been due to an unusually strong development of the warm Bering Sea ridge that winter. The paper also suggests that upper-air climate can be classified by characteristic appearance of graphs of the annual variation of temperature at each

level. For example, the Arctic and Antarctic have very different stratospheric patterns, which mere comparison of surface temperature range would not show. Also, the data showed that the layer up to 100 mb above the tropopause in the Arctic appears to be almost barotropic from 60° to 90°N with respect to mean annual temperatures, that the minimum annual range in temperature is almost constant at all stations, and that the actual value of the mean temperature at the level of minimum change is the same, about −53°C at all stations.

In 1954 the Arctic research group was transferred to McGill University from UCLA. In 1955 the University of Washington in Seattle started an Arctic research program. There the emphasis has been on analysis for the North Pacific and North America using the cross-section technique. The work done at McGill and Washington will be discussed in individual papers at this meeting. To my knowledge, no other organized program of Arctic meteorological research is being undertaken at any other university or research institute in the United States or Canada. However, there are several branches of the Government which are making some special effort toward better understanding of Arctic meteorological problems. The Canadian Meteorological Service, since 1950, has sponsored a small group under the direction of Mr. H. Wilson, at the Edmonton forecast center. It now prepares two surface and two 500 mb circumpolar maps daily south to about 50°N, on a slightly delayed basis, so that there is time to receive all late teletype reports and to be able to utilize the synoptic development on the detailed maps for North America which are also prepared to meet the ordinary operational requirements at Edmonton.

Suppl.
Part. I
1956

Among the government services in the United States, the U.S.A.F. Air Weather Service has several large forecast centrals and is able to investigate certain problems of immediate interest for its operational needs. Thus, at the centrals and major stations in Alaska and in Greenland, local forecast studies and regional climatologies can be prepared. However, Air Force meteorological research is normally performed by the Geophysical Research Directorate of the Air Force Cambridge Research Center, which in turn supports many contracts with individual universities and research institutes, such as those mentioned at UCLA, McGill and Washington, to carry out applied or basic research.

The United States Navy has had an Arctic area weather research program of its own to help it with the problems encountered in Arctic and Sub-Arctic waters. This program is under a general research project abbreviated as "AROWA" which started about 1951 and which has published forecast studies for the Bering Sea–North Pacific area, and for the Davis Strait–Canadian Archipelago waters. They have also selected typical weather patterns for the Arctic from the U.S. Northern Hemisphere Historical Map Series, reanalyzed the data for the area north of 50°N, and published surface and 500 mb charts as a guide for their forecasters. Mr. Wilson, of the Edmonton Arctic Forecast Group, assisted the AROWA project in these tasks.

As to Arctic meteorological research in other countries, I know of no group concentrating on such problems. However, a great deal has been done independently by interested meteorologists, especially in Germany, such as

the well-known research of Professors Flohn and Scherhag who will discuss some of their results tomorrow. The Soviet Union is active in all phases of Arctic research and no doubt has done considerable work in the meteorology of their sectors of the Arctic, but unfortunately any recent results, if published at all, are unknown in the West. A part of this lack of contact is due to the absence of any regular translation service to make Russian meteorological literature accessible.

III

We have seen so far that study of the aerological material for the Arctic, first available from about 1940 in Alaska and especially since 1946 from northern Greenland and northern Canada, has only recently begun. The basic problem upon which information is most needed is that of the nature of the general circulation in polar regions. To examine this directly from synoptic charts, one must have a reliable series of circumpolar maps for frequent levels from the surface to the highest level for which data can be taken, the checked and complete aerological data in published form, and summaries of the data, in order to digest them for comparative purposes. It may be useful to mention next the Arctic area charts and publications which are today available for this purpose, and to point out where improvement is needed.

Among the best-known circumpolar map series is that of the U.S. Weather Bureau's "Daily Series Northern Hemisphere Synoptic Weather Maps", familiarly called the "Northern Hemisphere Historical Series". The series starts with 1898 and presents 1 surface map daily to 1944. From then on, a 500 mb chart each day is included together with tabulated upper-air data for the time of the omitted 500 mb chart. The publication is usually released years after the map date. In recent years considerable effort is being made to obtain by mail from each country more complete and accurate data than possible from the teletype network. With respect to its Arctic analysis, the series is not sufficiently frequent, reliable, or complete, since the Arctic data are so often missing or garbled. The most recent years appear to have been prepared with much more care with respect to the Arctic, but I believe it is impossible to use any once-a-day series for any kind of sequential analysis in the Arctic.

A new research series which concentrates its efforts on the 50 and 100 mb levels is now appearing, which is also prepared by the U.S. Weather Bureau. It started with January 1953 and has so far published through May 1953. The series is made with all possible care and should provide a very useful tool for initial stratospheric analysis. Again, diurnal changes in pattern are not at all obvious and can only be guessed.

The German *Täglicher Wetterbericht* maps are well known. They are the best published maps I know for coverage of the Eastern Hemisphere. They, again, are published but once a day and are adequate perhaps for temperate latitude interpolation in the westerlies, but too infrequent for the cellular, irregular behaviour of Arctic centers. An excellent monthly summary of *Täglicher Wetterbericht* is issued as *Grosswetterlagen Mitteleuropas*.

The Edmonton series, already mentioned, is complementary in its coverage to the German map series, since it has the most dependable and complete

polar analysis available for the Western Hemisphere on a current basis. The series is used for research on the numerical specification of the pressure pattern problem at McGill, which will be discussed in a later lecture here. Recently the entire original Edmonton series has been microfilmed and is thus now available for research elsewhere.

There are, of course, many other current circumpolar analyses prepared by the weather services of other countries, but these are not published and are usually prepared for routine forecasts, under severe time limits which do not allow sufficient time for receipt of all data or for careful analysis, and thus are useful mainly for their frequency.

Of the many attempts to depict the mean monthly pressure pattern for the Arctic, the most recent and probably the most reliable, is that based on the 5 year mean of monthly *Grosswetterlagen Mitteleuropas* charts for 1948–52, prepared by Professor Scherhag's group in Berlin.

To sum up, it would be ideal if the advantages of each series could be collected to combine good coverage of both east and west, for several levels, with at least 12 hr continuity.

With respect to the publication of data, there are more serious deficiencies. Although surface data are summarized and published by almost every country in extreme detail, very few countries attempt to publish even summaries of their aerological material from which any future research progress must come. The intermittent series of upper-air data in *Daily Upper Air Bulletin* represented a weak attempt to provide this, but the Arctic data were very incomplete and unreliable, because of its duplication of unchecked teletype data. The new series of Northern Hemisphere Historical Map data which is being published daily on a 6 months' delay basis, starting with July 1955 data, inaugurated by Dr. Landsberg of the U.S. Weather Bureau, is the best development in regular presentation of data ever to appear. When other countries cooperate regularly so that its coverage is more inclusive, research requirements for actual observations should be mainly satisfied. If a second, monthly, publication which presented statistical summaries of the data could now be started, the data handicap would become a relatively minor, rather than the present major obstacle to current synoptic or climatic research.

IV

To illustrate the handicap of the data publication deficiency, the following two examples of research which could be done with existing but generally unavailable Arctic data, are mentioned. Although there are many investigations which might be undertaken, the following problems seem to me of great importance and they could be investigated if it were only possible to have the raw data processed. I wish to emphasize this point: additional observations will help, but we already have information which could yield untold riches for both the theoretical and practical meteorologist. What are really needed are processing facilities and accessibility to observations already stored and unused in our archives. When we consider the cost of taking the observations, indeed, of maintaining a weather service, or the cost of a single airplane accident due to weather which we cannot at present forecast, the cost of data processing and analysis appears insignificant.

As an example of a synoptic problem there is that of the polar trajectory of anticyclones. This involves a study of what appears at times to be anticyclogenesis in the polar areas; the origin, migration and development of coldest air masses; the more general problem of understanding just how an anticyclone "moves" or changes in intensity outside of the influence of the westerlies. Such a study would require frequent upper-air maps and detailed cross-sections based on verified data, for the entire circumpolar region since lobes of widespread high pressure areas extend in all directions.

Among the many climatological studies that can be made, possibly one of the most important and interesting is that of the polar stratospheric temperature field. We believe it is asymmetric with respect to the poles, but so far we do not have any reliable information concerning the degree of asymmetry of the mean monthly pattern, nor are we even sure of the general vertical thermal structure over the Arctic throughout the year. A good estimate of this should be possible simply from the summarizing of checked Arctic radiosonde data for the past 10 years. This same data could also give us our first information on other aspects of the temperature field, such as: How is insolational heating distributed through the atmosphere from polar night to polar day? At what levels, when, and with what intensity are these extreme changes accomplished? What are the thermal gradients and rates of change of the gradients in three dimensions? How do the temperature and wind fields in the stratosphere reverse with season? What is the nature of sudden temperature changes, or of the wandering of distinct warm and cold pools?

The Arctic is an ideal region to examine the thermal behaviour of the atmosphere since it is only in the polar areas that we have conditions of totally absent or continually present insolation. For this reason it would also seem a good area to look for those solar-terrestrial relationships which may be evidenced by temperature changes.

Thus, although more aerological data, especially from the Polar Basin, are needed, there already exist enough to give us some important information.

I will only conclude with the hope that we soon will be able to see the quick publication of individual and summarized aerological data in a standard form. This should apply especially to the data to be collected during the International Geophysical Year.

DISCUSSION, 2 JULY 1956

Chairman: J. NAMIAS, U.S.A.

NAMIAS: I think we have most interesting sessions in store for these next few days. For my own part I feel that we are at a stage in the study of the polar atmosphere which is most intriguing and interesting. As Dr. Sverdrup's remarks indicate, the most exciting and probably the most romantic periods of polar exploration have passed. These were in the days when it was really difficult to gather observations, and when it was not easy to interpret a few scanty observations in polar regions.

We have now come to the point where there is, relatively speaking, an abundance of data. Yet, we do not yet possess adequate information to nail down our subject, so that speculation is still necessary. It is my personal feeling that some degree of speculation, buttressed with facts as they gather over years of observation, make this phase of a science especially attractive, interesting, and of course exciting. By the time a science passes into the realm of complete objectivity or, in other words, the completely numerical stage, a lot of the life-blood has been extracted. In these next few days, as we see in the titles on our programme, we are in the happy intermediate phase where we can still talk and speculate; where we can debate, have controversies, and enjoy the skirmishes that most of us in science look forward to a great deal.

It is quite fitting that we open this symposium with a general treatment of the climate of the Arctic regions. It is particularly appropriate because, in order to study the irregular or non-seasonal variations, we must first have a background upon which to establish and assess them.

SVERDRUP: I would like to present a few remarks pertaining to the paper by Mr. Holcombe and point out certain differences between conditions in the Arctic and the Antarctic which I do believe have considerable bearing upon the various presentations of the general climatic contrasts. On one of the graphs the frequency of precipitation in the north and the south was shown, but there are hardly any data sufficient to show the contrast in the amount of precipitation. What I would like to remind you of is the fact that all over the Polar Sea the precipitation is very small. It may be a matter of between 100 or 150 mm, that is, from 4 to 6 in. annually, and a considerable part of that precipitation falls in summer. Similarly around the land areas immediately surrounding the Polar Sea precipitation is also small. One result of this is that during the early part of the Arctic summer the snow, which has collected on the polar ice, melts completely. Actually the snow-cover on the flat floes of polar ice is quite small because some of the snow has been carried towards the pressure ridges and has accumulated around these. The melting of the snow in early summer leads to a very great change in the albedo. During the winter the albedo is probably between 75 and 80. When the snow melts, the albedo decreases to perhaps 60–65, and most of the land areas surrounding the Polar Sea will be completely free from snow, and there we find an entirely different sub-surface in winter and in summer. In contrast, the precipitation around the coastal areas of the Antarctic is undoubtedly much higher. There do not exist very many measurements, but at the base of the Norwegian-British-Swedish Expedition 1949–52 the annual precipitation was of the order of 400 mm, about 16 in., and it is also striking that the ice surrounding the Antarctic continent, the pack-ice, receives a very heavy snow-cover, compared to that which is characteristic of the pack-ice in the north. The

Antarctic ice is even more uneven and broken up than is the pack-ice of the Arctic, there are fewer of the large flat floes and a thick snow-cover over all irregular ice. The result is that the snow-cover is so heavy that it does not disappear in summer, wherefore the belt of drift-ice retains its high albedo also during the Antarctic summer season. I am not quite sure that I can generalize as much as that, but on my single journey to the Antarctic I was struck by the fact that towards the end of the Antarctic summer I saw absolutely no snow-free ice in the vicinity of the continent. On the Antarctic continent itself we find a situation which is entirely different from that of the Arctic. The Antarctic continent is completely snow-covered, and the snow-cover is characterized by an exceedingly high albedo according to the measurements carried out by Mr. Liljequist. In winter the albedo may be as high as 90, and in summer it will be somewhat lower, but will hardly go below 80. Thus, in the Antarctic the basic characteristics of the sub-surface are more or less similar in winter and summer. As a consequence one should expect that the seasonal difference both as to average temperatures, pressure patterns, tracks of cyclones and so on, should be much greater in the Arctic than in the Antarctic. That of course does not mean that the circulation pattern in the Antarctic is simple. It appears that many irregularities exist, and I believe it will be a great advantage that during the Geophysical Year some stations are located fairly near each other. If one should have only stations far apart, then the observations of the Geophysical Year might also give the impression of a uniformity which does not exist, but that picture will be broken by a few stations close to each other. This is only an incidental remark. The principal feature that I would like to underline is the difference in the precipitation in the north and the south, particularly on the outskirts. Over the greater part of the Antarctic continent we have undoubtedly a very low precipitation. The precipitation at the South Pole has been estimated at only about 10 mm a year, if I remember correctly, but it is the precipitation at the outskirts which to a great extent will have a bearing on the relatively small differences existing between the summer and the winter months. That is the one point which I would like to present.

FRITZ: I wonder if I could ask Dr. Sverdrup a question. I was interested in the remarks about the low albedo in the Arctic. He said it was as low as 60–65 per cent. At T-3 albedo measurements were made in August 1955, but not at the part where they found the water over the ice itself; the albedo was 75 per cent. Would you consider that as much too high; or was it a non-representative value?

SVERDRUP: Yes, certainly I would consider that as too high a value when applied to the entire Polar Sea. It should be remembered that in the Polar Sea numerous small pools of water are found on a considerable number of ice floes.

BATES: First commenting on this precipitation problem, it is true that in the Polar Basin the ice melts at about a foot, about 350 mm, a year. Last summer we had a case of new snow falling on old snow on Ice Island T-3 and also off the coast of Alaska. There was a highly developed polar vortex. Some people called it, "the year summer did not come to the Arctic Basin". In contrast, of course, I think most of you have seen pictures of Little America which has been in existence now for about nearly 30 years. It is already buried, I would estimate, about 40–50 ft, so that certainly the precipitation there is at a rate of close to 500 mm a year net gain.

SUTCLIFFE: I would like to refer to the rather great differences between the two régimes in the Northern and Southern Hemispheres. In the Northern Hemisphere our cyclones are deeper in the winter of course, and in the winter we have rather more intense anticyclones on mean maps over the Pole. Then in the summer the polar anticyclones are weaker and the cyclones also. In the Southern Hemisphere, I am told, we cannot find any evidence of a more intense winter anticyclone.

NAMIAS: I have always been fascinated by the comparison that Dr. Sutcliffe indicated—between circulation over the North Pole and the South Pole. In the

Northern Hemisphere we have a phenomenon called the index cycle, which is essentially a tremendous growth of the polar anticyclone at the expense of the subtropical anticyclones. Index cycles have a maximum frequency of occurrence around February and the later part of winter. What makes the pressure at the surface so high, I believe, is the presence of a very high and cold stratosphere superimposed upon a shallow cake of extremely cold air which occasionally drifts over from Siberia. In this manner the net weight of air from the surface upward can result in amazingly high pressures. This index cycle, as it is called, is well known and has been studied at great length in the Northern Hemisphere for more than 10 years. As far as I know, there has not been much study of the same phenomena in the Southern Hemisphere. A very interesting question arises as to whether similar processes take place in the Southern Hemisphere, and if so on what scale.

FLOHN: The only sources of information for comparing the high-level circulation of the Southern and Northern Hemispheres are some small samples of wind measurements in middle latitudes. From an investigation of these data I found that the frequency of easterly winds, say at 500 mb and higher, in the latitudes between 45–55° in the Northern Hemisphere varies around 25–30 per cent at different longitudes. But in two series of the Southern Hemisphere at the same latitudes the frequency of easterlies and other winds indicating a blocking situation is only 1 to 3 per cent. Regarding the annual course of blocking situations, the question arises: why do we observe in the Arctic a large frequency of upper-air anticyclogenesis in spring, especially in March and in April, and not during the winter? Dealing with rather reliable series of surface pressure, one finds always that the frequency of Arctic anticyclones has a maximum in spring, connected with a spring maximum of low-index situations in middle latitudes.

Suppl.
Part. I
1956

NAMIAS: These are the anticyclones that have their genesis and development mainly within the subpolar region. During index cycles the greatest anticyclogenesis frequently occurs directly over the Polar Basin or at times over the Greenland–Davis Strait area.

One of the problems important to dynamic meteorology is whether exchange mechanisms might be of a different scale and character in the Southern Hemisphere than in the Northern Hemisphere.

SCHERHAG: With respect to the question why we have the highest pressure in the polar region in March and April, it may be that in March and April the whole land area over Siberia becomes warmer and then only in March and April the polar region is the coldest part of the Northern Hemisphere. And I believe that only at this time of the year we have to expect there the highest pressure in the mean. This high pressure will be reduced from April to May and June and then we have the greatest pressure fall in the polar region.

FLOHN: I am not quite sure if this hypothesis of Professor Scherhag is quite satisfactory. The warming of the Siberian area begins in March at the ground, but aloft it is stronger between April and May.

SCHERHAG: We are now at our Institute constructing new normal maps showing the mean temperature within all layers up to the 100 mb surface. And from these maps we can clearly see that this warming in Siberia begins even in February. There is here a strong warming between March and April. The polar region is the only area showing no warming in this period.

BATES: The question I had specifically in mind is addressed to Professor Hare. I have heard some botanists say that the tree line is merely an indication of what the climate was, say, 50–100 years, or perhaps several centuries, ago; in other words, that the tree line is actually delayed climatically. I believe in the Aleutians trees are as much as 300 miles from where trees would actually grow if you physically took them out and planted them in advance. Does this agree with the latest thinking?

HARE: There is, of course, a great deal of evidence that the tree line is not in equilibrium with the contemporary climate. It is exceedingly difficult today to find in the north any vegetation which is ecologically stable. But my point for attaching weight to the tree line, and most specifically the southern limit of the forest tundra, is not the ecological importance of that line at all; it is that you go from one world into another physically, from a meteorological point of view, as you cross it. As you do so you enter a different roughness régime, you enter a different moisture régime, you enter a different albedo zone; in other words you change worlds from our point of view. I have never yet met two ecologists who agree on the question that you have just asked me. So I have no opinion, except that it is obvious to anyone who walks across any of these divides that they are unstable at the moment, and it may well be that they are engaged in movement which is in response to changes that took place some time ago; but, as I say, from the meteorologist's point of view I do not think it matters. The migration is so tiny that from our point of view far more weight should be attached to the physical significance of the line today.

A great deal of weight is attached in the study of climatic change to the spread of *Sphagnum* moss, and the resulting overwhelming of parts of the forest, in at least two parts of the European post-glacial climatic sequence. And this is normally ascribed to a lowering of the mean temperature and increase in precipitation, or generally speaking both. There is evidence that *Sphagnum* bogs in the Labrador peninsula have been spreading at the expense of the forests in the past 30 years, during which time there has been a marked upward swing of temperature and no evidence whatever of increase in precipitation. In other words, I do not believe that this spread of the *Sphagnum* bogs is in any simple relationship whatever with temperature and precipitation. And inferences drawn from what happened in Sub-Atlantic times on this basis are inherently dangerous.

SVERDRUP: In regard to this very last question it may be of interest here to add that a certain region of the central part of the Scandinavian peninsula, just on the border between Sweden and Norway, was examined very thoroughly some 30 years ago by one of our botanists, who just a couple of years ago revisited the region. This is a region which is relatively little affected by the climatic change which was demonstrated by Hesselberg. But it was quite evident that the forest had crept upwards and that actually the vegetation in the higher regions was considerably greater than it had been some 30 years ago, which would indicate that at the borders the vegetation responds fairly quickly to changes in climate.

HESSELBERG: When examining the effect of the climatic changes on the vegetation we found that in Norway the tree line had risen about 50 m in about 50–60 years. We asked people in all parts of the country. Also the tree line had gone further to the north by about 100 or 200 km.

NAMIAS: You spoke this morning, Dr. Hesselberg, about the climatic change in Norway and Spitsbergen. Do I interpret you correctly by saying that this period of climatic warming which took place earlier, probably beginning before the turn of the century, is now at an end and is either levelling off or going in reverse?

HESSELBERG: We find the same thing in Norway as at Spitsbergen, that temperatures after the war are decidedly a bit lower than they were before the war. If this is only a small scale fluctuation or the end of the climatic warming period, we cannot know until we know more about the whole mechanism of climatic changes.

FRITZ: The Jan Mayen curve showed that the temperature had dropped in the period up to about 1941 or 1945, and then it rose again from the 1940s up to recent time. This curve was based on observation. Dr. Hesselberg mentioned also that there was correlation between the temperatures at the various stations. So I wonder if these correlations indicate any warming also at Spitsbergen in the period when the

data there are missing, so that the temperature actually might have been rising at Spitsbergen in the last 10 years.

HESSELBERG: I do not think that it has been rising in the latter years. There are so good correlations between temperatures at Spitsbergen and Vardø, where we have observations from the wartime. The correlation between Jan Mayen and the other stations is very low, 0·2 or something like that. Jan Mayen does not belong to the same well-correlated region with regard to temperature as Norway and Spitsbergen. It may be that the change in circulation brings more northerly winds into the Jan Mayen region.

NAMIAS: I am interested to hear that, Dr. Hesselberg. I would like to show a slide indicating some changes in the circulation which may be associated with that arresting of the rise. This slide is part of the paper I am going to deliver tomorrow. If the rise were still going on, as the slide will show, we would have to introduce some mechanism other than advection to account for it. Many investigators of climatic fluctuations in the Scandinavian area have pointed out that advection plays a dominant role in bringing about the climatic warming.

My conclusion, after demonstrating for you this change in circulation, is that it is quite probable that the recent climatic fluctuation can be linked to the change in general circulation patterns.

HESSELBERG: The communication of Dr. Namias was very interesting to me. I may mention that a paper is just now in printing giving the climatic variations in the last 10 years. And they show exactly the same thing as you mentioned. There is in this cooling area a more northerly wind component. We find a very close connection always between the modification of the general flow and the temperature variations. But what is the cause of these variations? And there I think we have to concentrate our research now, and I do not think we can find the cause without taking into account the varying activity of the sun.

MITCHELL: I should like to ask Dr. Hesselberg about the extreme temperature variations which he indicated in his paper at Isfjord and Vardø, especially at Isfjord. There was a very remarkable warming which occurred there in the period around 1915–20, and I am wondering if you are satisfied that this change was surely a climatic shift or whether it is possible that there had been a change in the procedure or in location of measurement of temperature at those stations. Some researchers in climatic changes like to look at climatic changes as a matter of shifts from one steady state situation to another, and I do not see anything too wrong with supposing that such shifts in the climate could occur in the region you have discussed. But I did notice that the form of the really very large shift in temperature there could exactly have been explained by a shift in the exposure of the thermometer in those years.

HESSELBERG: There was no change at the station during those years from 1912 up to 1930. Because we were aware of the very great changes up there, we very carefully examined the observations in those years, and we found no reason not to trust them, so we must take them as real climatic changes. And I will mention again, that also in Northern Norway we had a rather quick rise of temperatures in just the same period of years. And further I may mention that the people up in Tromsö were very astonished. "It is no longer dangerous to go to the Arctic region, . . ." it is too good to be true. Can it last?" "No," I said, "it probably will not last"; but it did last.

FLOHN: The results which Dr. Hesselberg derived from Arctic stations in the Atlantic area are confirmed by the well-known records from the Greenland stations. In addition to this, I would be interested in similar evaluations from the Alaska records, and new evidence of the climatic changes in Alaska and the area of Bering Sea.

BATES: I have no temperature records, but I do know that there have been relatively fantastic changes during the past 40 years in glacial ice conditions of the bays and valleys in Alaska.

Suppl. Part. I 1956

HARE: The analyses carried out by Longley of a long period record in Eastern Canada show a parallel with Dr. Hesselberg's result, and they show similarly very rapid changes at about 1920. It is, however, difficult to see how one can explain these parallel changes over large areas by advection only.

SUTCLIFFE: It is generally assumed that this is an advective effect which can be obtained from the mean monthly or the mean seasonal isotherms and contour lines; is it not thought to be an effect of greater exchange, greater activity, greater turbulence, rather than steady advection? It could be both of course. The difficulty of steady advection is that warming up in some areas means cooling somewhere else.

If it is a greater exchange or turbulence effect, then it can be in the same sense all the way around, it does not matter what the direction of the local prevailing wind happens to be.

I am interested in this point; you were rather interested in it yourself, Dr. Namias, whether we really have gone over the crest, whether we have any confidence in this change in temperature trend during the last decade. I am told that at Isfjord Radio, during the climatic warming period, the mean wind increased quite noticeably. I am told that the mean wind force at Spitsbergen was 2·7 from 1931 to 1934, becoming only 2·0 from 1947 to 1953. I am told that the average number of calms was only 11 for a period 1936–38; since the war the number of calms has gone up to 66, 33, 31, 49, 32 in the recent years. I am told that at all stations over this area of Greenland, the western side and the eastern side as well, the frequency of high winds has quite clearly gone down since about 1938 or 1939. This is probably not due to changes in instruments or to new instructions issued from the headquarters.

I would then draw attention to the work at another of our projects, working with the mean temperature at sea level all over the Northern Hemisphere, or over a large portion of it. The curves certainly show a definite falling rate from about 1938–48.

I have always been puzzled about this variation depending on the sun, where it seems we are working with a fraction of 1 per cent of the incident energy. Thus with a variation in the boundary conditions in the order of 1–1000, the system itself is to respond with variations which are almost qualitative.

In the Northern Hemisphere the whole belt of cyclones may shift into another part for a whole winter or a whole summer. Some summers have many anticyclones, others more cyclones. My point is that if with a small disturbing influence a system is oscillating in a rather big way, it would almost certainly oscillate in much the same way without the small disturbance.

HESSELBERG: Up in the mountain, in central Norway, on Dovre, we also have this very large change in temperature. And that is certainly because we have more cyclones in winter taking away the very cold surface layer. I believe that is one of the most important causes for the temperature rise at Spitsbergen in the warming period.

SECTION II

Suppl.
Part.I
1956

Supp
Part
1956

THE GENERAL CIRCULATION OF THE LOWER TROPOSPHERE OVER ARCTIC REGIONS AND ITS RELATION TO THE CIRCULATION ELSEWHERE

JEROME NAMIAS
U.S. Weather Bureau, Washington, D.C.

Arctic, polar and subpolar data extracted from twice daily analyzed charts of sea-level pressure and 700 mb height have been averaged for January, April, July and November of the 8 years 1948–55, during which there has been an unprecedented increase in meteorological observations at high latitudes. The resulting mean charts are compared with earlier estimates of the general circulation in polar regions and particularly with more recent charts published by the U.S. Weather Bureau. While there are significant differences, they do not for the most part alter the large-scale aspects of the patterns presented in the recent estimates, nor for that matter the gross features of sea-level normals published by Baur almost 30 years ago.

The major differences of the 1948–55 mean charts from the earlier U.S. Weather Bureau normals appear to be:

1. *An increase of the order of a few millibars in sea-level and mid-tropospheric pressures in January, April and November over the Davis Strait-Baffinland–Greenland area.*
2. *A decrease in sea-level and upper-level pressures of 1–4 mb in the center of the Polar Basin for all 4 months analyzed.*
3. *A greater tendency during the non-summer months for the Aleutian Low to be split into two cells (one off Kamchatka and the other in the Gulf of Alaska), with the corresponding emergence of an additional trough in the planetary circulation.*

The differences in July are everywhere small, amounting to less than 2 mb at sea level and aloft, except for an area of 2–3 mb fall in the Polar Basin.

An attempt is made to decide whether these differences represent errors in the earlier normals or whether they are transitory climatic fluctuations. A study of the prevailing anomalous circulations elsewhere in the hemisphere when Davis Strait abnormalities are of the type described under item 1 above, suggests that the 8-year period may have been somewhat anomalous in the north-east Canadian area.

On the other hand the suggested diminution in pressure over central portions of the Polar Basin is believed to represent an error in the earlier estimates. It is accounted for by difficulty in detecting disturbances breaking away from the Aleutian, Icelandic and Davis Strait centers of action, especially during periods of pronounced blocking.

The net effect on the circumpolar zonal circulation north of 70°N is a suggested small increase (1 m/sec or less) in west wind drift aloft and a corresponding decrease of the surface easterlies.

Relationships between the Arctic and the complete hemispheric circulations are described, with special emphasis on the interactions attending the development of deep winter Arctic anticyclones and the apparently interdependent deep summer Arctic cyclones.

I. Introduction

The necessity of meteorological data over polar regions for purposes of weather analysis and prediction is now obvious even to the novice. The present generation of meteorologists is especially fortunate in being able to realize earlier hopes of obtaining a reasonably accurate synoptic picture of the three-dimensional structure of the atmosphere over the entire polar area. These data have brought to light many phenomena of circulation and weather hardly suspected in former years and thereby have been vital in the development of scientific long-range prediction. Indeed, students of long-range problems have contributed materially to knowledge of the polar atmosphere, for they have realized perhaps more than others the impossibility of segregating it from the remainder of the hemispheric circulation.

One of the principal problems at once arising in connection with synoptic or climatological studies of the Arctic is the normal state of the circulation at any time, for any irregular fluctuations of short or long duration must be evaluated against this background. It is not surprising, therefore, that considerable attention has been devoted to this question throughout the past 50 years, at first with data only for sea level, and later with information from the upper air.

The history of attempts to piece together normal patterns of circulation over the Arctic may be broadly classified into 5 stages:

1. A period involving speculation based largely upon theoretical reasoning and from circulations in other latitudes. Maury and Ferrel, for example, would be among the list of pioneers in this endeavor.
2. A period involving the formulation of quasi-empirical models of circulation based upon fragmentary information gathered with great effort and even peril during expeditions such as that of the *Fram* and *Maud*.
3. The phase wherein investigators attempted to utilize peripheral and expedition data in a "synoptic" fashion—that is, by plotting averages on maps and then attempting judicious interpolations and extrapolations of isobars. Especially noteworthy in this phase is the work of Baur[1], Dzerdzerskii[2] and Shaw.[3]
4. The period where upper-level normal patterns were attempted largely by employing extrapolation techniques from sea-level constructions in addition to some peripheral upper-air averages. For example, the early work of Shaw[3], Byers and Starr[4] and Scherhag[5] belongs to this stage.
5. The current effort involves the processing and synthesis of vast amounts of surface and upper-air data obtained mainly in the past 15 years during which there has been an unprecedented increase. Examples may be found in studies carried on in Great Britain under the direction of C. E. P. Brooks[6] and in the U.S.A. of the author of this paper.[7]

The present note synthesizes some newer observations gathered from the past 8 years, compares these with earlier estimates, and discusses briefly the general problem of interdependence of the Arctic, temperate and subtropical circulations.

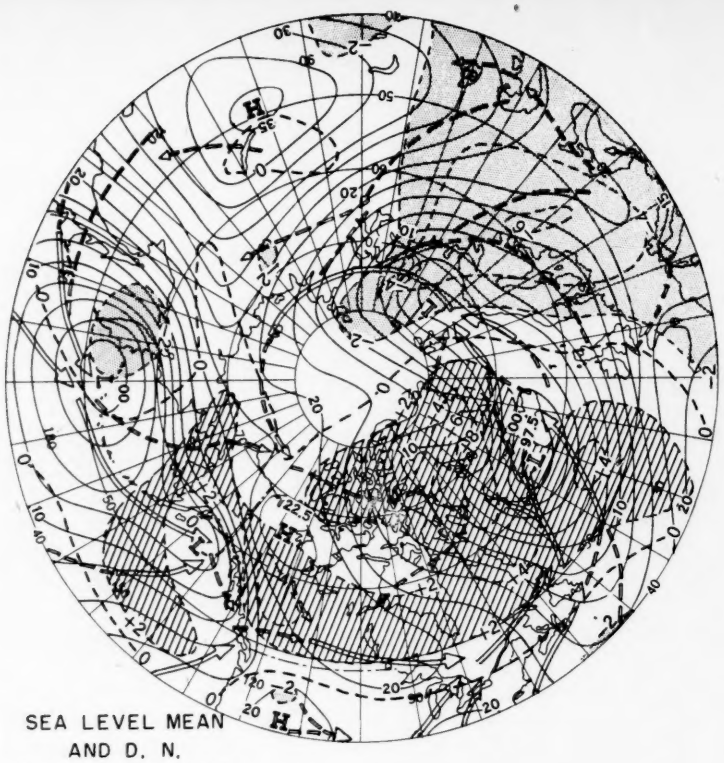
II. Comments on the Synoptic Approach

The construction of a 40 year series of daily Northern Hemisphere map analyses, one of the most ambitious meteorological undertakings during World War II, would appear to be an ideal source of information from which to determine normal patterns of sea-level circulation. Shortly after the completion of these charts values of pressure were interpolated from the isobars for a closely spaced grid of points and 40 year averages for each month were computed and published in the form of isobaric charts.[8] While these charts probably gave a better picture of the normal sea-level circulation over many portions of the Northern Hemisphere than previously available, they did not improve the picture over Arctic regions. In fact Rodewald,[9] comparing these polar and Arctic pressures with those given by Baur[1] and by Sverdrup,[10] and with 1949's pressure distribution, concluded that the 40 year averages were appreciably too high over the entire Arctic. A revision of the normals, using more refined methods of preparation and data up to 1950, verified Rodewald's conclusion which indeed had been suspected by forecasters even at the time of computation of the 40 year averages.

This paradox wherein a systematic analysis and synthesis of an increased supply of data can lead to *poorer* estimates of a normal or mean picture deserves further discussion, for it unveils a trap into which the purely statistically-minded meteorologist is apt to fall. Besides, such further discussion is appropriate since the newer charts presented in this report have also been constructed in a statistical fashion by averaging interpolations from a grid of latitude-longitude intersections on twice-daily charts.

If the scale of the circulation phenomena analyzed is smaller than the distance between observing stations, as it was over polar regions in practically all 40 years of the historical map series, many systems (cyclones and anticyclones) are missed and the analyst draws a picture which he thinks is characteristic for the particular area. Now the preparation of the historical map series, a war-time high-priority project, had to draw upon large numbers of young meteorologists, many of whom could hardly be called mature analysts. It should not come as a surprise that these men frequently constructed great anticyclones over the Arctic Basin, for in their university courses they were frequently taught about the steady and characteristic anticyclones there. Thus averages computed from interpolations of their isobars would naturally reflect this bias. On the other hand, analyses constructed from mean values computed at peripheral stations and from a few expeditions, in the hands of one experienced with the scales of these mean systems, could be interpreted more correctly. The dimensions of mean or time-averaged systems is appreciably greater than that of the day-to-day cyclones and anticyclones, a fact well known to workers in the field of "centers of action" even at the turn of the century. Hence, in 1929 Baur was able to construct sea-level maps for the Arctic which are in better agreement with our newest estimates than were the 40 year means.

When the grid of stations becomes relatively dense so that the day-to-day cyclones and anticyclones are easily detected and defined, the grid-averaging system is indeed good and is perhaps the best one to obtain normals. One of its virtues is that it contains automatic smoothing and detection of errors by the very nature of synoptic analysis. In other words, when a point is reached



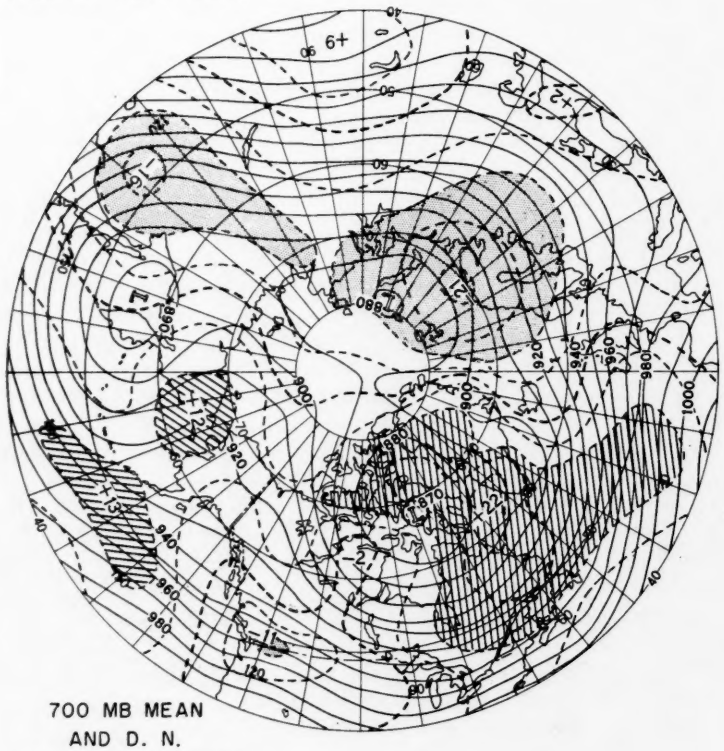
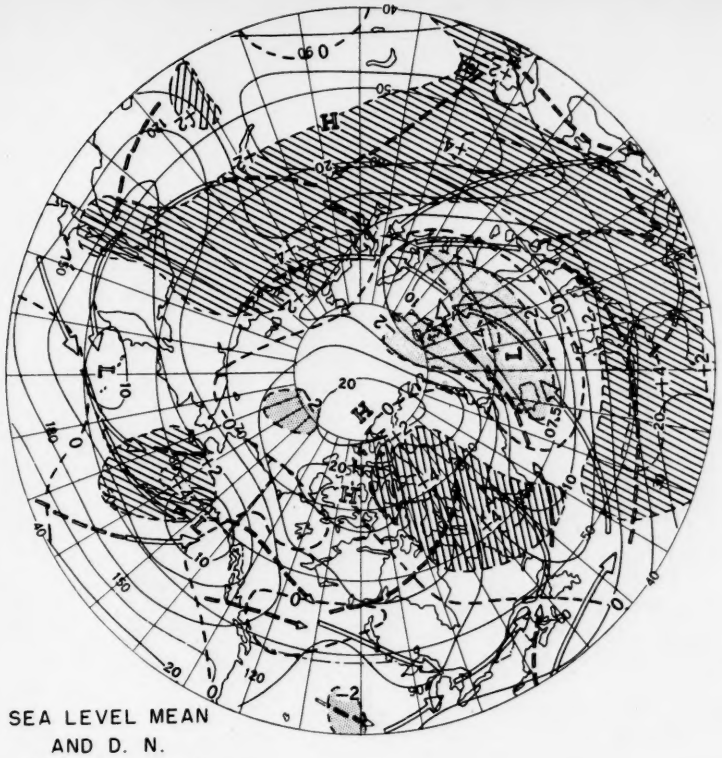


Fig. 1. Mean sea-level isobars (above) and 700 mb contours (below) for the months of January 1948–55. Broken lines give changes from earlier U.S. Weather Bureau normal estimate[7]. Further details given in text

Suppl.
Part. I
1956



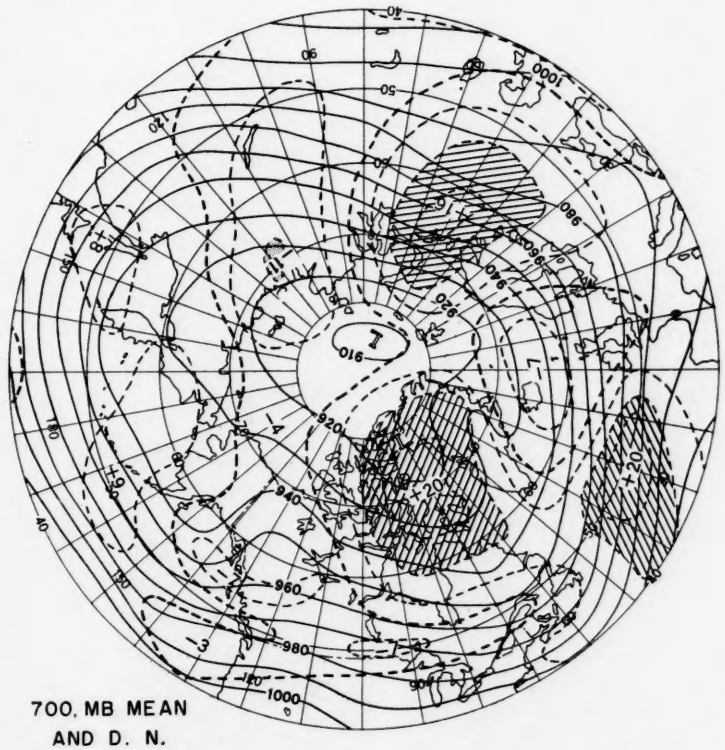
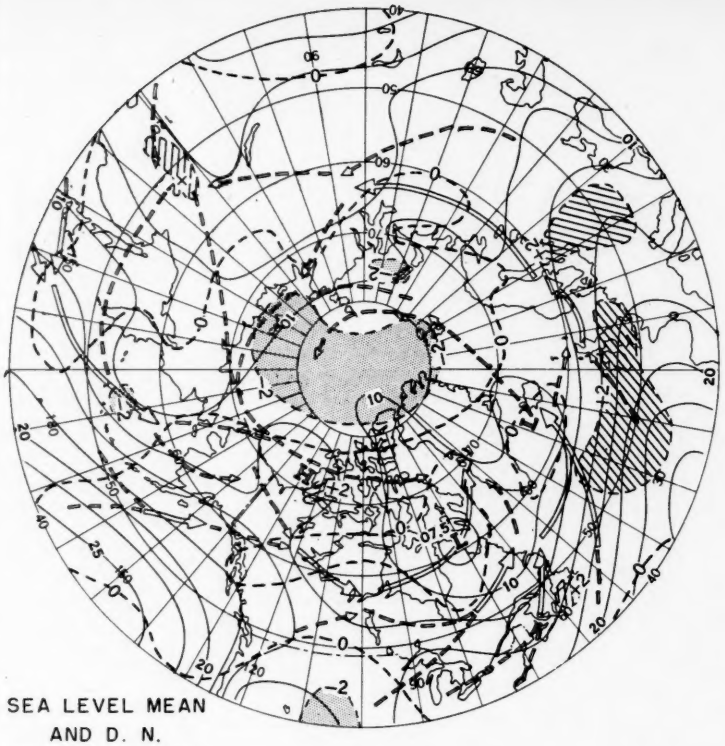
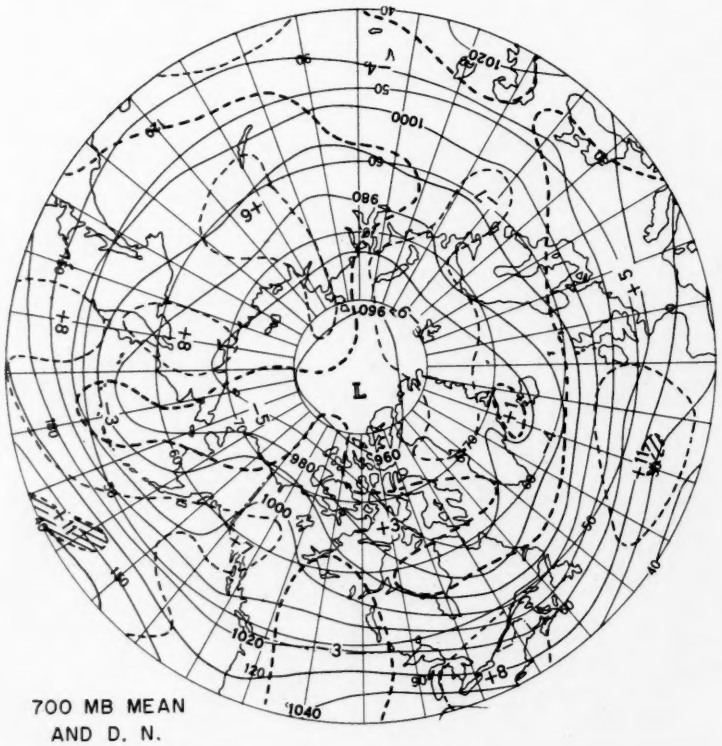


Fig. 2. Same as Fig. 1 but for April



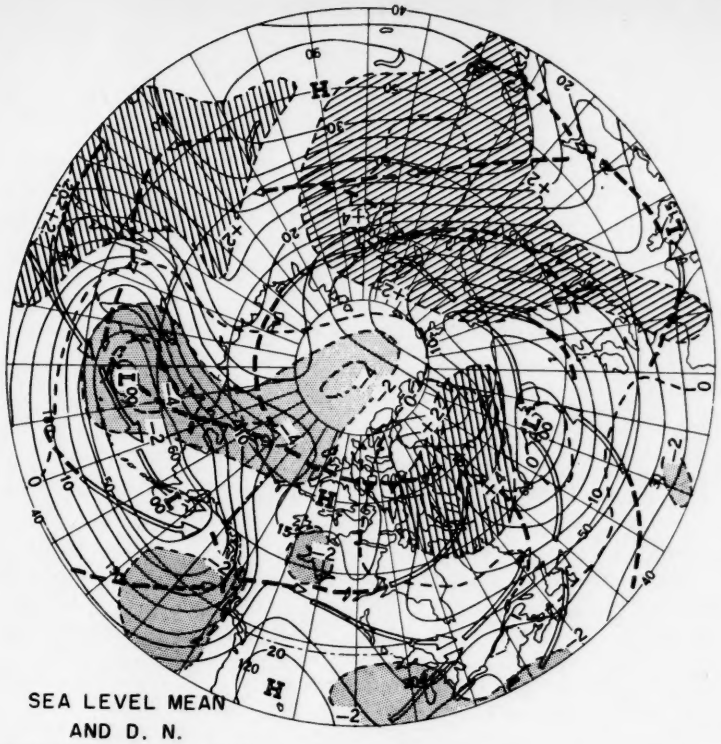
SEA LEVEL MEAN
AND D. N.
JULY 1948 – 1955



700 MB MEAN
AND D. N.
JULY 1948 – 1955

Fig. 3. Same as Fig. 1 but for July

Suppl.
Part. I
1956



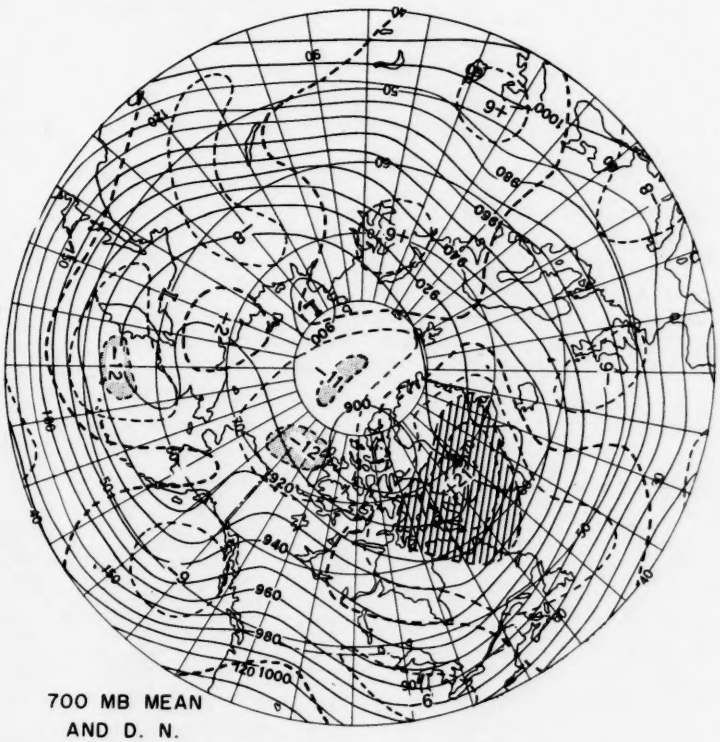


Fig. 4. Same as Fig. 1 but for November

where the density of the grid permits delineation of scale phenomena of the size of cyclones and anticyclones the grid-averaging system becomes effective.

III. New Data and Comparisons with Earlier Estimates

As mentioned earlier, an extensive attempt to obtain better estimates of the normal monthly pressure distribution at sea level and in mid-troposphere was made from 1950–52, based upon data up to 1950.[7] For Arctic and adjacent areas the procedure laid considerable stress on data from the five years 1946–50 when Arctic observations increased materially. The actual procedure (detailed in the publication[7]) employed a weighting system whereby the longer series of data along a roughly elliptical area enclosing the Arctic were judiciously blended with the newer Arctic material derived from 5 years of twice-daily synoptic charts. A number of tests for internal consistency were applied to these charts so that upon their completion there was a general feeling of satisfaction. On the occasion of the AGARD Polar Atmosphere Symposium a new series of charts for January, April, July and November of the 8 year period 1948–55 have been constructed in order to see if there were substantial deviations from the earlier charts, and if so, to suggest modifications. These mean charts were prepared from grid-point averages taken from twice-daily routine hemispheric analyses at sea level and 700 mb. Since 1948 these analyses over Arctic areas have been reasonably complete and homogeneous. The results are shown in *Figs. 1* to *4*. Sea-level isobars are drawn for each $2\frac{1}{2}$ mb and 700 mb contours for each 100 ft. The grid of points at which intersection pressures or heights were read was sufficiently dense to justify the indicated refinement of pattern—namely, for each 10 degrees of longitude at every 5 degrees of latitude up to 75°N; each 20 degrees of longitude at 80°N and then the value at the pole.

Single broken lines in *Figs. 1–4* give the changes from the earlier Weather Bureau normals (published in 1952)—isopleths are drawn for every 2 mb change at sea level and every 100 ft at the 700 mb surface. Hatched areas represent *increases* greater than 2 mb at sea level and 100 ft at 700 mb, while dotted areas show *decreases* of the same order. On the sea-level charts "normal storm paths" taken from work by Klein[11] are entered as open solid (primary paths) and broken (secondary paths).

Although a lengthy description of the circulation features on these charts could be given it is felt that the user can readily extract from them what information he wishes without this description. Moreover, the essential large-scale features of the charts do not depart radically from the earlier estimates, descriptions of which may be found in many published texts and papers. Expressed in other words, it appears that the differences between earlier and recent estimates are small enough relative to the observed gradients that the form of the circulation over polar and Arctic areas is much the same. The area enclosed by differences of 4 mb at sea level and 200 ft at 700 mb is indeed small, and over many areas (i.e. the unhatched or unstippled areas where the changes are less than 2 mb at sea level and less than 100 ft aloft) rather small differences result in the basic patterns expressed by the two estimates. It is not meant to imply here that the changes are insignificant and that the earlier estimates are near perfect, but rather that our present

knowledge of the form of the general circulation of the lower troposphere is fairly satisfactory for certain large-scale studies. For example, it seems adequate for studies involving fluctuations of the kind experienced in 5 day or monthly means, which are usually many times the magnitude of the changes indicated in *Figs. 1–4*. In fact, as we shall see when examining the individual area deviations more closely, it is difficult to decide whether they represent errors in the earlier estimates of the normal or whether they are merely transitory climatic fluctuations in pressure pattern associated with the eight-year period.

In January (*Fig. 1*) the greatest deviations appear in the Davis Strait–Greenland area and adjacent portions (+ 8 mb and + 220 ft) and in northern Europe (−6 mb and −210 ft). A stronger Bering Sea ridge appears both at surface and aloft than on the earlier estimates and the Aleutian low is split into two cells (off Kamchatka and in the Gulf of Alaska) rather than one which appeared in the earlier January normal. The corresponding emergence of an additional trough in the planetary wave train occurs off western Northern America. A similar tendency to split cells also appears in connection with the Icelandic low, and at upper levels the Novaya Zemlya trough actually assumes the form of a closed vortex. In short, the new January chart is somewhat more meridional in character at high latitudes with the possible associated emergence of a new trough in the Gulf of Alaska and accentuation of the Novaya Zemlya trough.

Suppl.
Part. I
1956

In April (*Fig. 2*) the Davis Strait increase once more shows up but the companion decrease appears in the vicinity of Iceland rather than north Central Europe as it did in January. The Aleutian low again has a decidedly split structure compared with the earlier charts and again the Gulf of Alaska trough emerges.

In July (*Fig. 3*) the changes are everywhere small and the main shaded area (>2 mb at sea level) appears near the pole. In part the smallness of the summertime changes are due to the smaller variability of pressures at this season.

In November (*Fig. 4*) similar changes to January occur—particularly in the Davis Strait–Greenland area and again in the character of the Aleutian low.

Whether the deviations described above suggest revision of the earlier normals or are merely transitory climatic fluctuations characteristic for the period of averaging cannot be conclusively answered. However, there are some bits of indirect evidence which may throw some light. In the first place, large increases in mid-tropospheric pressure observed in the Davis Strait–Greenland area during winter are frequently associated with decreases in North and Central Europe. In other words the main change centers over North America and Europe may be companion phenomena. In fact, studies of the interdependence of mid-troposphere anomalies in different parts of the hemisphere have played an important role in extended forecasting practice for many years,[12] and MARTIN[13] has completed an extensive file of charts like those shown in *Figs. 5* and *6* illustrating these "teleconnections". Essentially, Martin selected for one area (like that given by the heavy black circle in the Davis Strait in *Fig. 5*) times when large anomalous centers were present, and then computed from a longer series of 5-day mean charts the

probability of sign of the anomaly in other adjacent and remote areas. Isolines of these probabilities were then drawn. This pattern can also be interpreted qualitatively as a pattern of probable anomalies attending the occurrence of the large deviations in the selected area. Comparing Martin's chart (*Fig. 5*) when a large positive anomaly is observed in the Davis Strait with the change pattern in the January 8-year mean (*Fig. 1b*), the similarity is striking, not only in the European area to the east but also in western

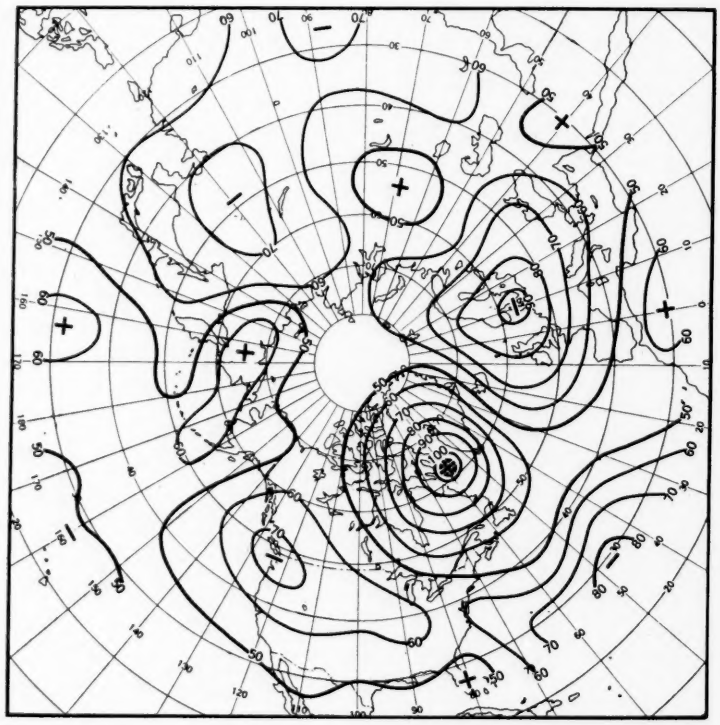


Fig. 5. Isopleths of probability of sign of mid-tropospheric anomaly (in per cent) when a positive centre of anomaly lies in the black-crossed (Davis Strait) area in winter. (See text for details of chart construction)

North America, the Bering Sea and parts of Siberia. In view of this correspondence, and in view of the fact that the length of record and thus reliability of the 700 mb normals is greater in areas removed from the Davis Strait, it would seem probable that the 8 year period (1948–55) has been anomalously high in this general area (Davis Strait–Baffinland–Greenland), and that the earlier normal pattern should not be modified much—certainly not to the degree indicated by the January changes. Unfortunately Martin's work could be completed only for the summer and winter seasons so that comparisons of this sort cannot be made for all the other charts presented. However, the chart selected for positive anomalies in the Davis Strait area in summer (*Fig. 6*) shows that the companion negative center is

apt to be found not far from Iceland, rather than in Europe, again suggesting that the increases over north-east Canada to Greenland in April (*Fig. 2b*) may be transitory.

The indication that the Aleutian sub-polar low normally consists of two cells, one near Kamchatka and the other in the Gulf of Alaska, is more difficult to assess. The fact that this tendency shows up on all charts except July suggests that this may be characteristic of the cooler seasons of the year.

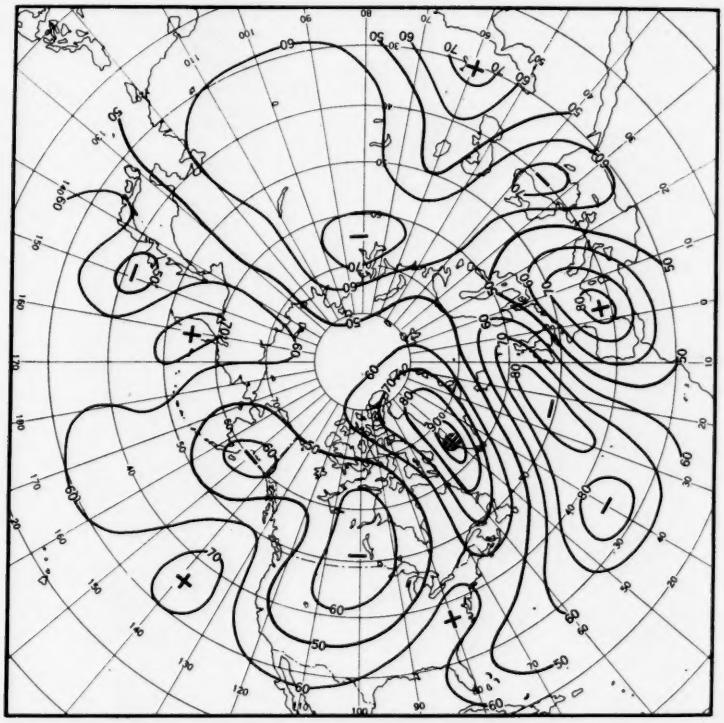


Fig. 6. Same as Fig. 5, but for summer

A further indication that the double low-pressure cell is more representative of the long-term normal in the Pacific than the single cell is that the enriched data of the last 8 years would facilitate analytical detection of secondary cyclones forming in the Gulf of Alaska while the primary occluded cyclone remains behind in the central or western Aleutians. The weak upper-level trough in the Gulf of Alaska is of course associated with the surface low-pressure center, and whether it is a transitory or permanent climatic feature is the same question.

When the circumpolar changes are averaged according to latitude (*Fig. 7*) it becomes clear that the zonal gradient from 70°N to the Pole has increased at sea level and aloft in all 4 months examined. Part of this increase is accounted for by decreases in the Polar Basin and part by increases at 70°N, some of which have been described. The consistency of falls in central

portions of the Polar Basin during all seasons and levels would suggest that the earlier normals were slightly too high, and that the "anticyclonic bias" of analysts has not yet been completely dispelled. Experience suggests that

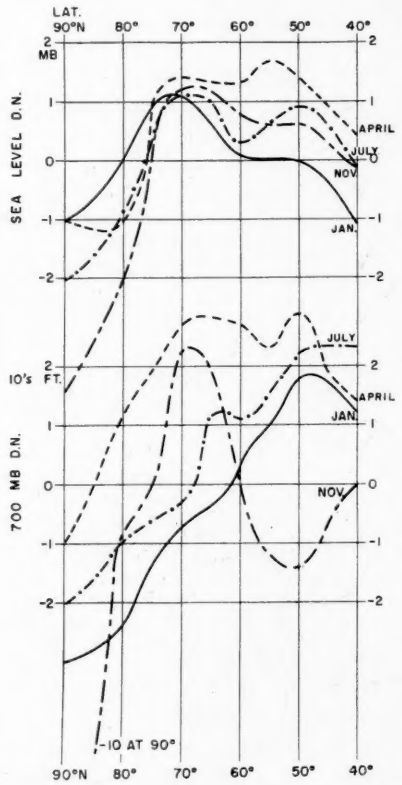


Fig. 7. Latitudinally averaged changes in sea-level pressure (above) and 700 mb height (below) from earlier U.S. Weather Bureau normal estimates[7] to the new 1948–55 averages

analysts have a tendency to neglect the formation of breakaway depressions which frequently develop out of Aleutian, Icelandic and Davis Strait cyclones, particularly during periods of pronounced blocking. Indeed these breakaway depressions are frequent enough to be represented as primary or secondary storm tracks on many of Klein's "normal storm track" charts (arrows in *Figs. 1–4*).

The net effect of the suggested increase in zonal flow surrounding the pole is, however, of the order of less than 1 m/sec which, for example, increases the 700 mb westerlies from their previous January normal of about 0·5 m/sec (from 70° to 90°N) to about 1·0 m/sec, and changes the strength of the normal east wind drift in January from about 2·0 m/sec to 1·5 m/sec. However, the well-known assymetry of the polar circulation[14] makes such comparisons have only restricted meaning.

Fig. 8 shows a comparison of latitudinally averaged pressures at sea level

as computed from the two sets of charts and also from Baur's (1925) values. The suggested relaxation of the east to west drift around the pole coupled with lower pressures there (than in the 1952 estimates) is clearly shown. Baur's estimates, considering the date of their preparation (1929), seem surprisingly good.

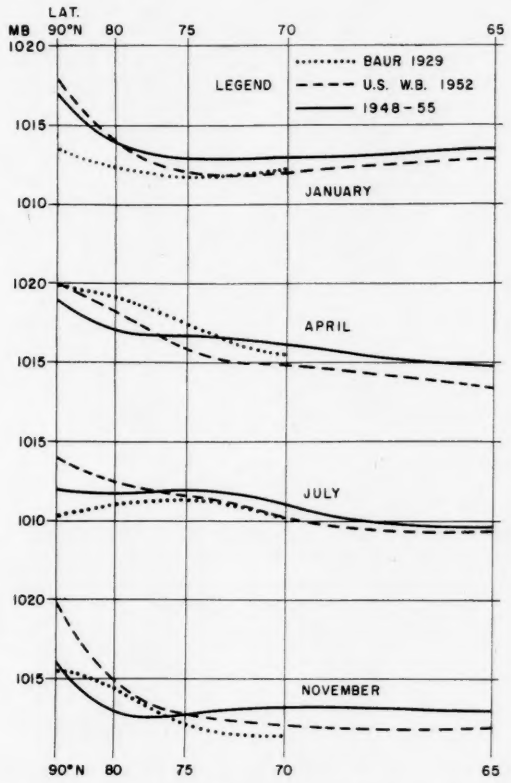


Fig. 8. Sea-level pressure profiles from 3 sources (see text)

IV. Relations of the Arctic Circulation to that Elsewhere

Inter-relations between the subpolar cyclonic centers of action and the subtropical anticyclones were recognized many years ago. For example, Walker[15] pointed out the negative correlations between Icelandic and Azores pressures, and also between pressures in the Aleutians and Hawaii. More recently the general see-saw between circumpolar pressures at 70° and 30°N was demonstrated by Lorenz.[16] These interconnections and many others of a regional nature often are associated with "index cycles"[17 18]—periods lasting from three to six weeks during which the circumpolar upper-level vortex expands and the Arctic and polar anticyclones grow while the subtropical highs decline, followed by a return to normalcy of the general circulation. A particularly pronounced cycle of this kind in 1947 strongly influenced the weather of the Northern Hemisphere during most of February.[19]

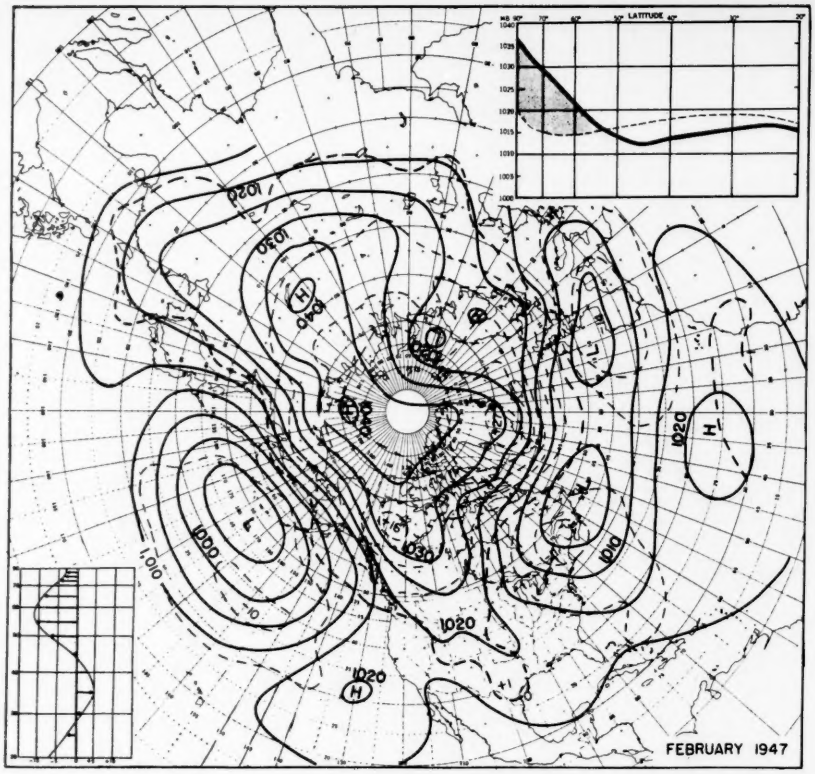
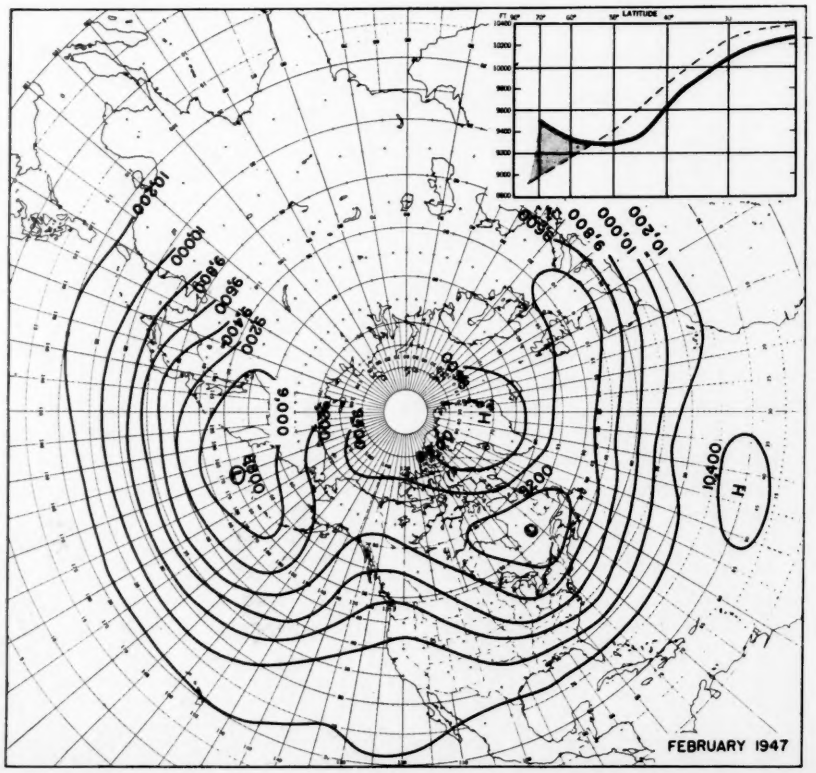


Fig. 9. (Above) Mean sea-level pressure and its anomaly (broken lines) for February 1947 and (below) corresponding mean 700 mb chart. Inserts in upper right give profiles alongside the normal (broken). In lower left of upper figure is shown the sea-level geostrophic wind component as a function of latitude



The great abnormalities of this month's circulation may be seen in *Fig. 9.* Note particularly the positive abnormality of pressures at high latitudes (profiles in upper right) and the compensatory negative departures at low latitudes. Note also the lateral (equatorward to 45°N) and vertical extent of the polar easterlies during this period. Even the average seasonal picture for the 1946–47 winter at the 700 mb level reflected the great southward expansion of the circumpolar westerlies.

Although these index cycles have been described and discussed at great length their physics remain obscure and only qualitative "explanations" of their origin and characteristics have been advanced. Consequently the important questions of why, when or just how such a cycle will proceed are as yet unanswered—leaving a great gap in extended forecasting procedures.

However, Martin's charts of simultaneous teleconnections between anomalies in distant areas frequently offer clues of what may lie in the Arctic area if the anomalies in other areas are known (see, for example, Martin's charts described in Dr. Fletcher's article in this volume).

Suppl.
Part. I
1956

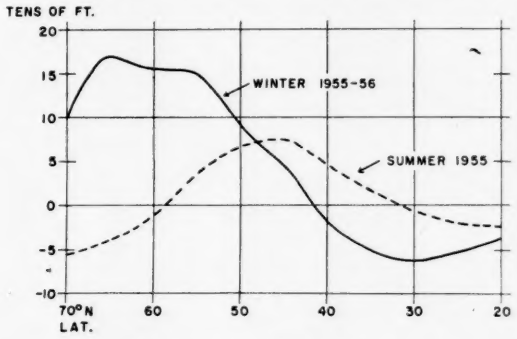


Fig. 10. Departures from normal of the 700 mb height profiles for the area 0° westward to 180°

While the index cycles described above are primarily cold season phenomena, there are apparently great and persistent abnormalities of the hemispheric circulation which also occur in summer. One of these aberrations which has a strong influence on the seasonal characteristics of much of the Northern Hemisphere occurs when a girdle of persistent positive anomalies appears in mid-latitudes, and is generally associated with a northward displacement of the subtropical anticyclones. Such a case is illustrated by the 700 mb anomaly profile for the summer of 1955 (*Fig. 10*), the regional constitution of which is shown in the map of *Fig. 11.* As in the summer of 1955, this type of northward displacement of the subtropical highs is apt to be associated with much stronger than normal westerlies (and increased cyclonic action) along the rim of the Polar Basin, and also with poleward displaced subtropical easterlies. The latter displacement usually leads to increased easterly wave and tropical cyclone activity. Because of the upper wind abnormality tropical cyclones formed in such a region are apt to pursue further westward courses than normal before recurving into the westerlies.*

* Further details of the circulation and weather of the summer of 1955 may be found in appropriate issues of the *Monthly Weather Review.*

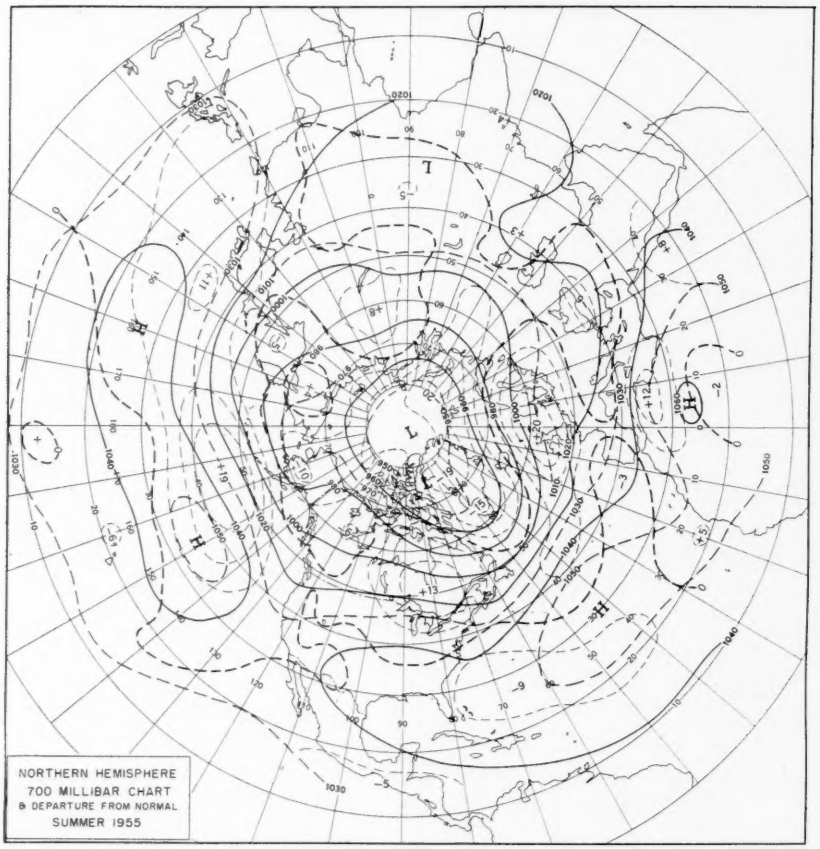


Fig. 11. Mean 700 mb contours and anomalies for summer 1955

Some recent work[20] indicates that there may be a connection between summers characterized by circulations of this anomalous type and the adjacent winters. That is, winters of greatly enhanced polar anticylcones (expanded circumpolar vortex like winter 1955–56 shown in *Fig. 10*) may tend to be embedded between summers when the circumpolar anomalies are of the type illustrated in *Fig. 10*. Such periods would be characterized by accentuated continentality at high middle latitudes because of the diminished winter westerlies and summer prevalence of upper-level anticyclones.

References

[1] Baur, F. "Das Klima der bisher erforschten Teile der Arkis" *Arktis 2*, p. 77, 1929

[2] Dzerdzerskii, B. L. "Tsirkuliatsionnye skhemy v troposfere Tsentral'noi Arktiki" *Moskva. Izdatel'stro Akademii Nauk* (*USSR*), 1945. English translation as U.C.L.A. Dept. of Met. Scientific Report No. 3, June 1954, under U.S.A.F. Contract 19(122)–228

[3] SHAW, N. *Manual of Meteorology* Vol. II—Comparative Meteorology, 2nd Ed.; Cambridge University Press, 1936

[4] BYERS, H. and STARR, V. "The Circulation of the Atmosphere in High Latitude during Winter" *Mon. Weather Rev. Washington*, Supplement No. 47, 1941

[5] SCHERHAG, R. *Neue Methoden der Wetteranalyse und Wetterprognose;* Berlin, Gøttingen, Heidelberg, pp. 424, 1948

[6] BROOKS, C. E. P., *et al.* "Upper Winds over the World" *Geophys. Mem., Lond. No. 85*, pp. 145, 1950

[7] U.S. WEATHER BUREAU. *Normal Weather Charts for the Northern Hemisphere.* Technical Paper, No. 21, pp. 74, 1952

[8] U.S. WEATHER BUREAU. *Normal Weather Maps, Northern Hemisphere Sea Level Pressure;* Washington D.C., 1946

[9] RODEWALD, M. "Zur Fragen Luftdruckverhältnisse in der Arktis" *Ann. Met.* Heft 9/10, 1950

[10] SVERDRUP, H. U. "Übersicht über das Klima des Polarmeeres und des Kanadischen Archipels" *Hand b.d. Klimatologie von Koppen-Geiger;* Bd. II, Teil K, I. Berlin, 1935

[11] KLEIN, W. H. "Normal Storm Tracks of the Northern Hemisphere" *U.S. Weather Bureau, Research Paper, No.* 40, p. 60, 1957.

[12] NAMIAS, J. "Thirty-Day Forecasting: a Review of a 10 year Experiment" *Met. Monographs* Vol. 2, No. 6*;* Amer. Met. Soc., Boston, Mass., 1953

[13] MARTIN, D. E. "Anomalies in the Northern Hemisphere 700 mb 5 day Mean Circulation Patterns" *Air Weather Service Rep.*, No. 105–100; HQ, AWS, Andrews AFB, 1953

[14] LA SEUR, N. E. "On the Asymmetry of the Middle-latitude Circumpolar Current" *J. Met.* Vol. 11, pp. 43–47, 1954

[15] WALKER, G. T. and BLISS, E. W. "World Weather IV—Some Applications to Seasonal Foreshadowing" *Mem. R. met. Soc*, Vol. 3, pp. 81–95, 1930

[16] LORENZ, E. N, "Seasonal and Irregular Variations of the Northern Hemisphere Sea-Level Pressure Profile" *J. Met.* Vol. 8, pp. 52–59, 1951

[17] ROSSBY, C. G. and WILLETT, H. C. "The Circulation of the Upper Troposphere and Lower Stratosphere" *Science*, Vol. 108, pp. 643–652, 1948

[18] NAMIAS, J. "The Index Cycle and its Role in the General Circulation" *J. Met.* Vol. 7, pp. 130–139

[19] NAMIAS, J. "Characteristics of the General Circulation over the Northern Hemisphere during the Abnormal Winter 1946–47" *Mon. Weather Rev.* Vol. 75(8), pp. 145–152, 1947

[20] NAMIAS, J. "Characteristics of Cold Winters and Warm Summers over Scandinavia related to the General Circulation" *J. Met.* Vol. 14, pp. 235–250, 1957.

Suppl.
Part. I
1956

RECENT STUDIES ON THE ARCTIC TROPOSPHERE AND ITS TELECONNECTIONS

H. FLOHN and G. SEIDEL
Frankfurt-on-Main

The investigations are based on the evaluation of the daily mean tropospheric temperature, represented by the thickness 500–1000 mb, over the Central Arctic (75–90°N) during each winter from 1949–50 to 1955–56 and of daily values of the area covered by Arctic air (defined through a thickness of less than 5000 geopot. m) as a whole and divided into four sections of the Northern Hemisphere. These data are compared with circulation indices of middle latitudes and particularly over Europe, since hemispherical indices are only of restricted value for regional forecasting. There was no evidence for regular and persistent periodicities in the Central Arctic. The average patterns of the general circulation in Sub-Arctic and Arctic regions are discussed.

The averaged tropospheric temperature of the Central Arctic between November and March shows strong irregular time variations. Direct relationship between the hemispheric zonal index and the mean temperature of the Arctic, as suggested by F. Defant, could not be verificated. The area of Arctic air over the North American sector is rather closely connected with the temperature of the Central Arctic; an increase (outbreak) of Arctic air in that section shortly precedes general warming of the Central Arctic.

Starting from the example of the cold period of February 1956, in Central Europe, a connection between the area of Arctic air over the Siberian section and the average zonal wind component at the 500 mb level over Europe (with a delay of about one week) could be found in several winters. Large-scale anomalies of the middle latitude circulation, especially the blocking of the westerlies, appear to be connected with asymmetric shifting of the Arctic cold air to one side of the polar region.

Additional statistical studies on the aerology of the cold air center in eastern Siberia confirm earlier results. The low variability of temperature above the inversion is discussed.

1. INTRODUCTION

EXPERIENCE of about 15 years medium-range forecasting shows that the concept of hemispherical zonal indices does not meet the needs of the forecaster dealing with areas of the order 10^5–10^6 km² like Central Europe. The ultra-large-scale teleconnections of the general atmospheric circulations do not present univocal solutions for regional forecasts, and it has been shown[1] that quite different large-scale weather patterns (Großwetterlagen) over Europe do not affect the hemispherical meridional mass exchange (Austausch) at 55°N latitude. Therefore it seems to be advisable, partly following some suggestions of H. RIEHL, F. DEFANT[2] and A. SIMILÄ,[3] to investigate the behaviour of the Arctic cap of tropospheric cold air in different sections, and their influence on distant regions. Since each winter should be regarded as an individual one (cf. *Fig. 2*), it is inevitable to extend this study to a comparison of several years, starting from 1949–50, when the aerological network was established sufficiently. Some preliminary results of this investigation shall be presented here.

Suppl.
Part. I
1956

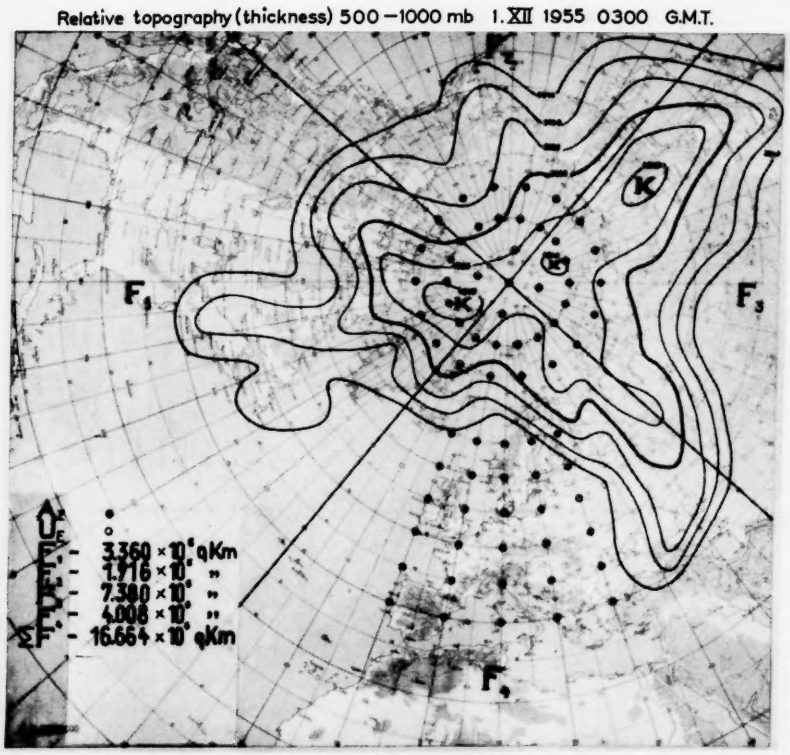


Fig. 1. Typical example of a daily map with the area determining A_z (each dot representing a grid-point) and the areas F_1–F_4 of the longitudinal sections (thickness < 5000 gpm)

(*Facing page* 62)

Supp
Part
1956

The evaluation of the data refers (*Fig. 1*) to the average thickness 500–1000 mb of the central Arctic north of 75°N (A_z) and the area (F) covered by an "Arctic air-mass" with a thickness of less than 5000 gpm, divided into four longitudinal sections (F_1–F_4). A comparison of the daily values of A_z

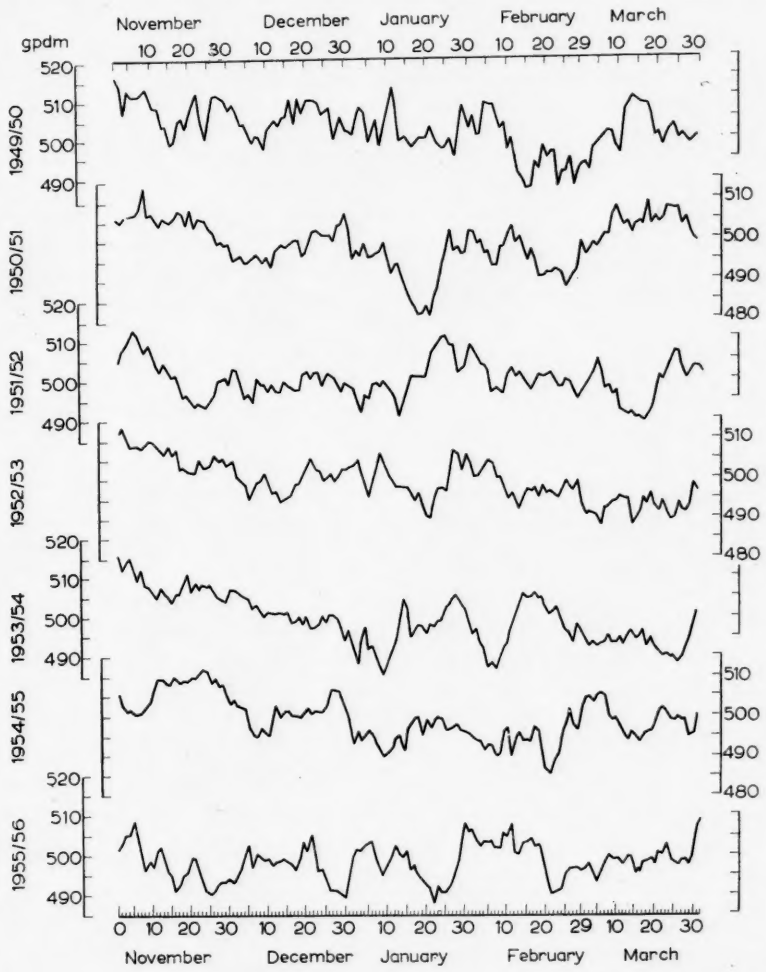


Fig. 2. Daily values of the average thickness 500/1000 mb north or 75°N (A_z) for each winter 1949–50 to 1955–56

for seven winters (*Fig. 2*) shows that the large-scale fluctuations of tropospheric temperature between mid-November and mid-March cannot be interpreted as a consequence of the seasonal changes of radiation balance in polar latitudes. These fluctuations appear to be quite irregular, and no persistent periodicity can be easily detected. A systematic comparison with daily values of the geostrophic wind zonal component at the 500 mb level,

at the area 40–65°N, 10°W–30°E (U_E, cf. *Fig. 1*)), and of a hemispherical zonal index U_{55} between 50° and 60°N[1] could not confirm the hypothesis[2] of a regular balance between the temperature of the central Arctic and the strength of the middle-latitude westerlies.

2. Permanent Features of the General Circulation

Our knowledge of Arctic meteorology depends to a large extent on the existence of regular ascents in the central part of the Arctic. The radiosonde stations on ice islands and—to a smaller degree—reconnaissance flights have substantially improved the reliability of our daily weather maps. If we compare the annual 500 mb map for 1949–51[4] with that of 1952–55 (*Fig. 3*), and regard the cold pole centers of individual months and years, we observe the ellipticity of the circumpolar vortex. Its core is stretching from Baffinland to Cap Chelyuskin; the annual centers are situated in 1949–52 near the magnetic pole, at about 76°N, but since 1953 we find these centers in close vicinity of the geographical pole, near 87°N, mostly on the Canadian side of the Arctic. It can be assumed that earlier results in this respect are biased by the lack of information. Some other features, e.g. the asymmetric extension of the Siberian trough to the Ochotsk Sea, and the well-known existence of winter centers in eastern Siberia, are to be seen in both maps.

The persistence of the anomalies of the circumpolar vortex can be seen from a sequence of charts for the months January, April, July and October (*Fig. 4*), representing the deviations from latitudinal averages north of 50°N. The annual persistence of the American trough contrasts strikingly with the variations of all other anomalies, especially in months of transition. We observe also a splitting of the Pacific trough during the warmer season. If we draw these deviations during the year in form of isopleths for a given latitude (*Fig. 5*), there is hardly a predominating feature at 80°N, while at 65° and 50°N the annual persistence of the North American trough is predominating. The other troughs are migrating and partly vanishing during the course of the year.

3. Teleconnections of the Arctic Air Cap in Winter-time

The basic question of long- and medium-range forecasting is to forecast the behaviour of individual and of time-averaged troughs and ridges and their deviations from normal. So we are obliged to study comparatively the teleconnections for individual winters, at first with rather primitive means. During winter 1955–56 some remarkable features occurred; at first the marked meridional temperature difference and the resulting strong zonal circulation over Europe from November to January, accompanied by a widespread coldness over North America up to December, and secondly, the persisting blocking anticyclones over north-west Siberia and Scandinavia during February and March, accompanied as usual by an intense period of coldness over Europe. The question arose, whether the large-scale dislocations of the Arctic cold cap are related to the fluctuations of the principal jet-stream zones and to the occurrence of persistent blocking anticyclones, as suggested already by Brezowsky.[5] The comparison (*Fig. 6*) of the sections

Suppl.
Part. I
1956

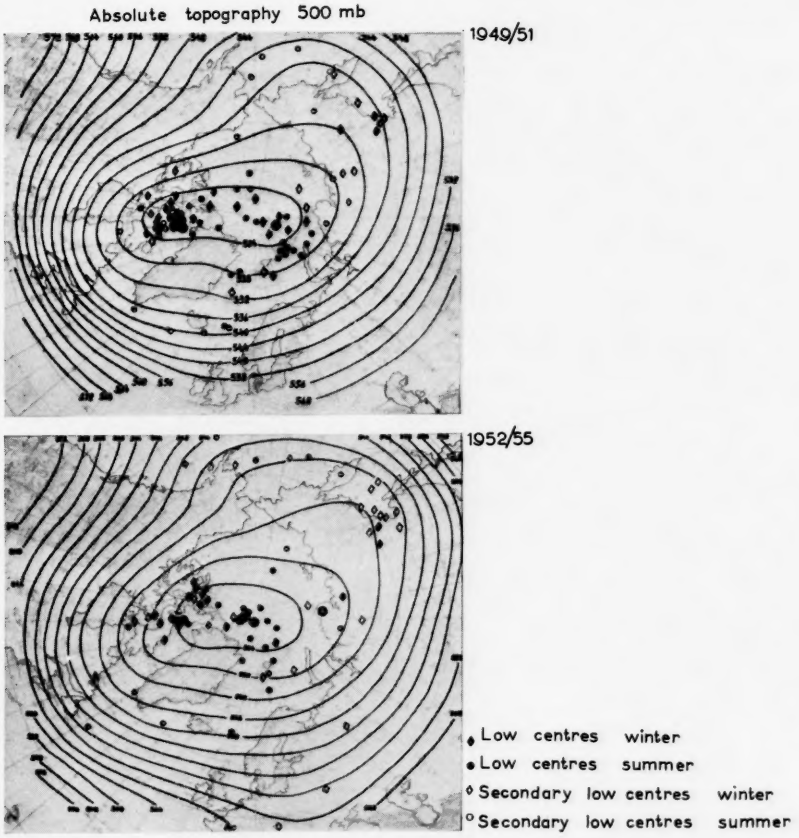


Fig. 3. Annual contour map of the 500 mb level, 1952–55, compared with 1949–51. Each dot represents a vortex center of an individual summer month, each quadrangle the same during winter, each circled dot a vortex core of an individual year

(*Facing page* 64)

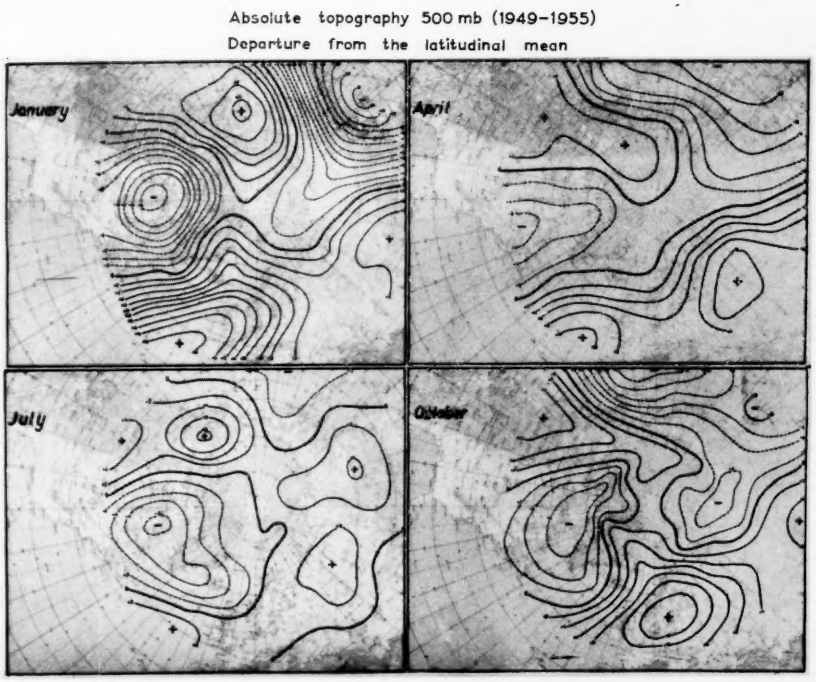


Fig. 4. Deviation of the 500 mb contour from latitudinal averages for characteristic months (averaged 1949–55), unity 10 gpm

(Facing page 65)

F_1 (America) and F_2 (Pacific) shows an inverse trend, which cannot be found by comparing F_3 (Siberia) with F_4 (Europe and eastern Atlantic). If we express the areas $F_1 \ldots F_4$ as percentages of F (*Fig. 6*) we observe that the sum F_1+F_2 contains not more than 40 per cent on most days till the end of

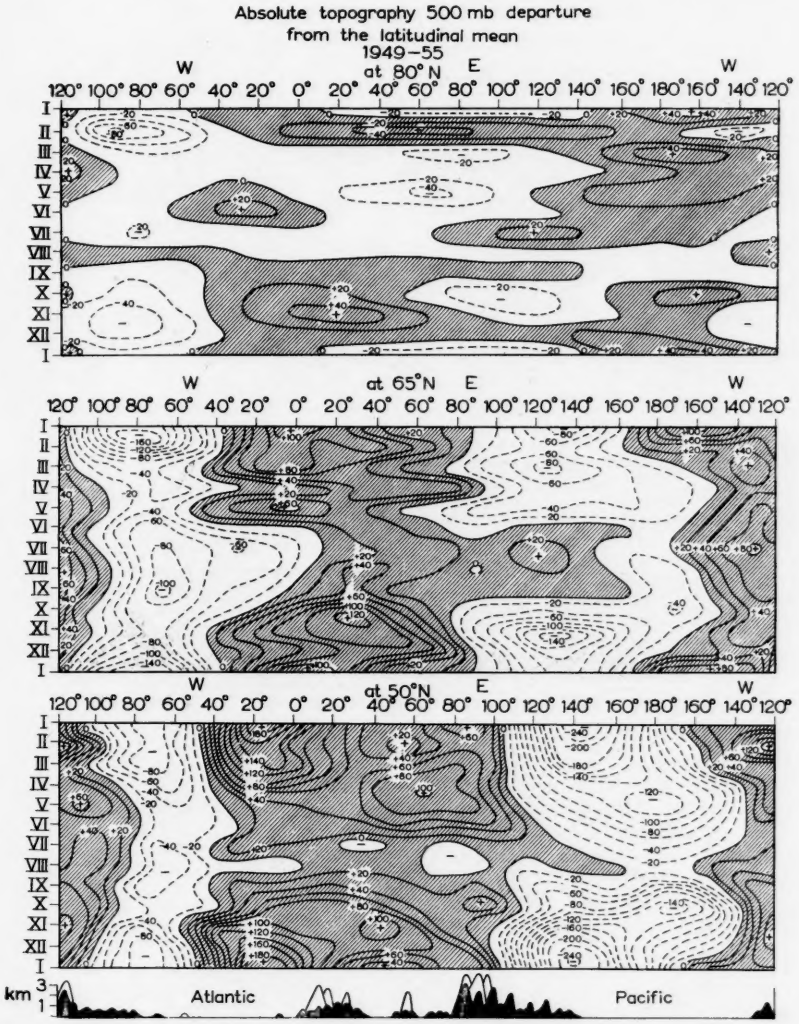

Absolute topography 500 mb departure
from the latitudinal mean
1949–55
at 80° N
at 65° N
at 50° N
W
E
120° 100° 80° 60° 40° 20° 0° 20° 40° 60° 80° 100° 120° 140° 160° 180° 160° 140° 120°
I II III IV V VI VII VIII IX X XI XII I
km 3 1
Atlantic
Pacific

Fig. 5. Isopleths of the deviation from latitudinal averages of 500 mb contour during the year at lat. 80°N, 65°N and 50°N, unity 10 gpm

January, but varies about 60 per cent during February and March. Obviously the cold air caps have been partly dislocated from the Siberian to the Canadian side of the Arctic. This happens together with a reversal of the zonal circulation over Europe and western Siberia.

Suppl.
Part.I
1956

By comparing the variations of the zonal component U_E with the extension of the cold air over Siberia (F_3) the maxima (or minima) of U_E follow about 5 days after a maximum (or minimum) of the area U_E with a time lag varying between 2 and 7 days (*Table 1*).

Although this relationship can hardly be understood immediately, we observe that high percentage values of F_1+F_2—which are represented also by small

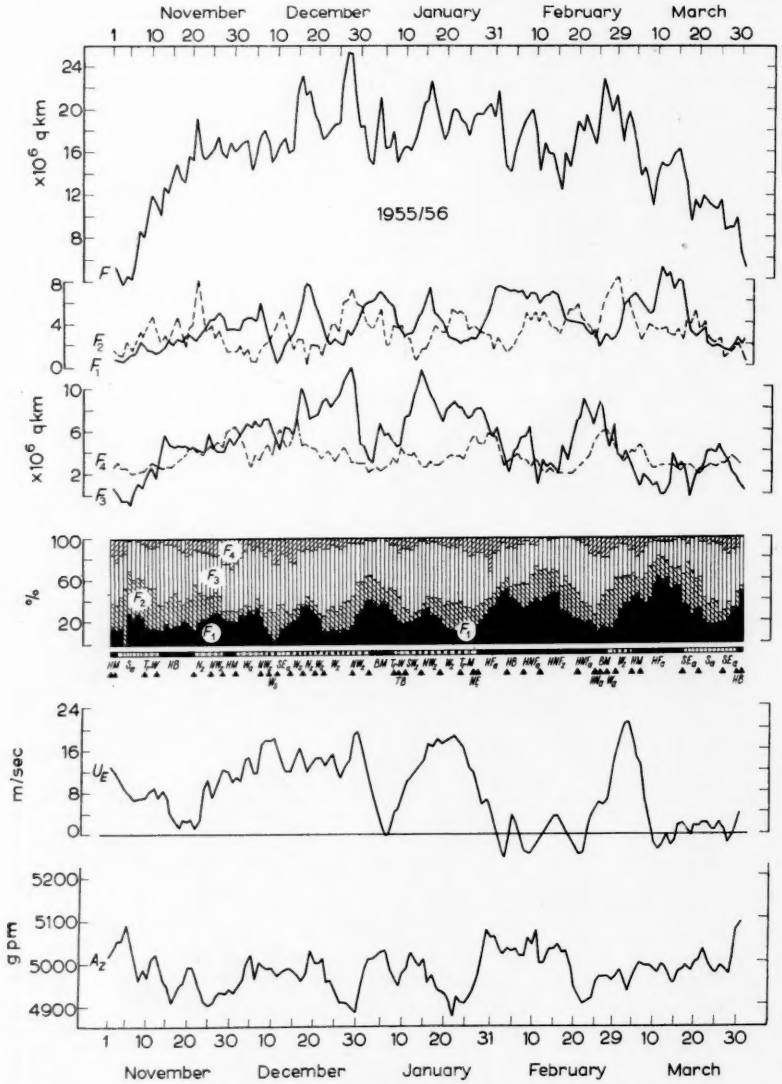


Fig. 6. Daily values of F (hemispheric area of thickness 500/1000 mb <5000 gpm), the same for Sections $F_1 \ldots F_4$, $F_1 \ldots F_4$ as percentages of F, U_E (area of Fig. 1) and A_Z during the winter 1955–56. In the center block the numbers at the per cent areas must be reversed: $F_1 = F_4$, $F_2 = F_3$, etc., below the European Grosswetterlagen

Table 1. Correlation Coefficients r between U_E and F_3 with Varying Time Lag $t(U_E) - t(F_3) = \Delta t$; n = number of pairs

Period (F_3)	Δt	$r(U_E, F_3)$	*n*
14.12.1950–22.2.1951	6	+0·67	71
10.1.1954–15.3.1954	2	+0·46	65
5.1.1955–24.3.1955	7	+0·33	79
15.12.1955–25.3.1956	6	+0·66	102

values of F_3, while F_4 is unimportant—are followed by blocking anticyclones over Europe, in correspondence to the results found by Brezowsky. So we may suggest the hypothesis, that large-scale dislocation of the cold air cap to the American section of the Arctic precedes the occurrence of persistent blocking over Europe and north-west Siberia by a few days. And vice-versa a strong development of the isolated cold air centre in Eastern Siberia (F_3) seems to be related with periods of high zonal index over Europe. Both suggestions will be checked systematically.

The fluctuations of temperature in the central Arctic (A_z) are rather closely connected with the extension of F_1, of the North American trough, as already suggested by F. DEFANT.[2]

Suppl.
Part. I
1956

Table 2. Correlation Coefficients r between A_z and F_1 with Constant Time Lag $t(A_z) - t(F_1) = \Delta t$; n = number of pairs

Winter (F_1)	Δt (days)	$r(A_z, F_1)$	*n*
5.1.1951–3.3.1951	2	0·60	58
1.1.1954–26.3.1954	2	0·35	85
1.12.1954–12.3.1955	2	0·29	100
1.12.1955–8.3.1956	2	0·54	99

Table 2 shows that with a time-lag of no more than two days the correlation between A_z and F_1 is positive: a cold air outflow over North America precedes shortly a warming of the central Arctic. Since a similar relationship could not be found over Siberia, the Siberian winter cold pole, which is situated very far from the pole, near 62° latitude, seems to be partly independent of the larger cold air centre fluctuating over the central Arctic and Canada.

Table 3. Correlation Coefficients between F_1 and F_3 with a Time Lag $\Delta t = t(F_1) - t(F_3) = 23$ resp 24 days; n = number of pairs

Period (F_1)	Δt	$r(F_1, F_3)$	*n*
15.12.1950–5.3.1951	24	0·82	81
15.12.1952–31.3.1953	24	0·59	107
15.1.1955–31.3.1955	23	0·44	76
15.1.1956–31.3.1956	23	0·32	77

The most surprising result could be found in comparing the curves for F_1 (America) and F_3 (Siberia) with a time-lag of 23–24^d, which cannot be easily understood from a physical point of view. But since this result is derived from four individual winters, and furthermore it is confirmed by rather high correlations (*Table 3*), it should be considered seriously.

It must be admitted, that the results shown in *Tables 1–3* are of a purely empirical nature, and that they cannot be shown in each individual winter. But since the need for long- and medium-range forecasting is so urgent, we feel it necessary to discuss such items, even if a sound physical explanation cannot be given at the present time. A more complete study of this subject has been recently published,[7] including graphs illustrating *Tables 1–3.*

4. Contribution to the Knowledge of the Siberian Cold Pole

In former investigations[4] it was found that over eastern Siberia and over the Canadian Archipelago the temperatures above the inversion change very little with the temperatures near the surface. Since this fact is not in conformity with experiences in middle latitudes, and contrasts also with calculations of radiational cooling, it seems necessary to check it with more recent data. From 9 selected radiosonde stations distributed over eastern Siberia (100–135°E), divided into 2 groups between 60° and 65°N ("S") and between 67° and 72° ("N"), we separated 4 samples of ascents (December 1955–February 1956) with respect to surface temperature (*Fig. 7*). This may be allowed, as the averaged temperatures aloft above the individual stations vary less than 1–2°C in this uniform climatic region. While the averaged surface temperatures of those classes vary between −19° and −49°C, we found at 500 mb only a difference between extreme classes of 3·7° at the S-group, 6·6° at the N-group, and at 300 mb—frequently above above the tropopause—this difference decreases to about 2°C. The standard deviation of temperature decreases rapidly with height (*Fig. 8*), while in

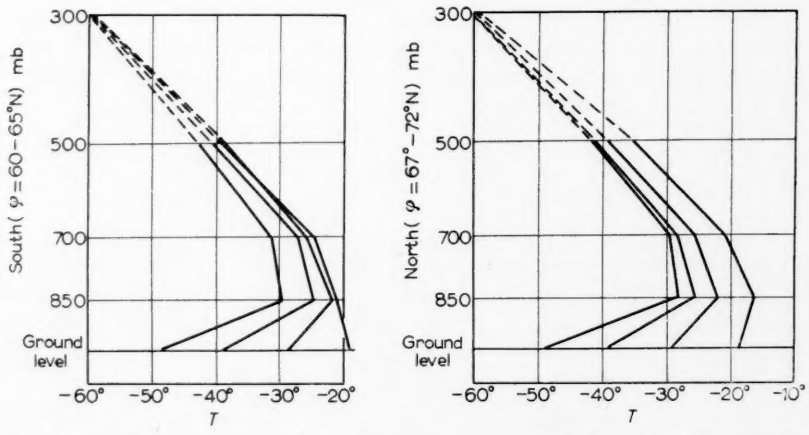


Fig. 7. Averaged vertical temperature distribution over Eastern Siberia (N = 67–72°N, S = 60–65°N) for samples separated with respect to surface temperature (December 1955–February 1956)

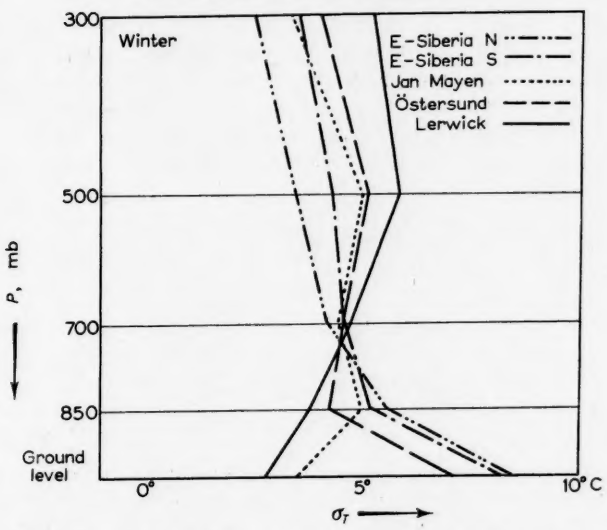


Fig. 8. Vertical distribution of standard deviation of temperature σ_T in different climates (winter)

maritime subpolar climates (cf. Jan Mayen, Lerwick) the largest deviations are to be found near the level of non-divergence.

Separating the ascents with respect to surface pressure (*Table 4b*), it can be shown that with low-pressure values as well as with extremely high-pressure values the average temperature is a few degrees higher than at medium pressures (1020–1039 mb). This may be interpreted as a result of warm air advection during cyclonic periods, or of increased subsidence in strong anticyclones. The results of a similar separation of group "S" with regard to wind measurements at the same levels (*Table 4a*) are not very conclusive due to the scarcity of winds aloft. But since the measurements are not substantially biased by the very low cloudiness, we can derive that at the 850 and 700 mb level north-westerly winds from the central Arctic are the coldest (and the most frequent), while easterly winds, pointing to the occurrence of a blocking anticyclonic cell over the northern coast, are 3–5°C warmer. At 500 mb the winds are rather uniformly distributed; those from

Table 4. Temperatures Aloft over Eastern Siberia (60–65°N, 100–130°E) as a function of (a) Wind Direction dd at the Same Level, (b) Surface Pressure PPP. (—) = number of cases, °C

(a)	*dd* =	*34–03*	*04–09*	*10–15*	*16–21*	*22–27*	*28–33*
	500 mb	−38·1(17)	−38·0(17)	−39·4(12)	−41·7(16)	−40·7(11)	−40·1(23)
	700 mb	−27·9(32)	−25·6(24)	−27·9(13)	−27·5(17)	−28·6(26)	−28·9(45)
	850 mb	−23·2(45)	−21·4(27)	−22·9(18)	−25·3(24)	−25·3(27)	−26·7(72)

(b)	*PPP* =	*>1019*	*1020–29*	*1030–39*	*<1040 mb*
	300 mb	−58·1(36)	−59·8(21)	−58·7(23)	−58·8(43)
	500 mb	−39·7(89)	−41·2(110)	−40·3(108)	−38·4(103)
	700 mb	−22·5(91)	−28·9(111)	−28·0(110)	−25·3(103)
	850 mb	−22·9(89)	−26·4(111)	−25·7(108)	−22·5(101)

south—just preceding a marked trough—are 3–4° colder than those from north to east, occurring at the rear of a trough or during blocking situations.

Some tentative calculations deal with the magnitude of subsidence in anticyclonic cells. The relative humidity of the Siberian winter ascents (*Table 5*) at 500 and 700 mb is apparently higher than that over Eureka and Thule (48–57 per cent, cf.[4] *Table 18*), but sources of instrumental error cannot be excluded.

Table 5. Average Difference between Temperature and Dew-point (T-Td) and Relative Humidities (RH); n = number of cases

	T-Td			*RH*			*n*		
	850	*700*	*500*	*850*	*700*	*500*	*850*	*700*	*500 mb*
Jakutsk	3·7	4·1	3·7°C	71	68	68%	83	79	64
Shigansk	4·2	3·9	3·7	67	69	68	74	76	62
Werchojansk	3·6	3·2	.	72	74	.	48	43	.

If we select a sample of persistent anticyclonic cells, we can calculate the average temperature change during 24^h at the same station; then we may estimate the magnitude of subsidence by means of the adiabatic equation with the same method as used by H. FAUST.[6] The results (*Table 6*) are not convincing, but they seem to reveal a very small magnitude of subsidence, as well as *Table 5*.

Table 6. Average Interdiurnal Temperature Changes (ΔT_{24}) and Subsidence (w) in Anticyclones

		850	*700*	*500 mb*	*Zero-layer*
ΔT_{24}	Europe	+1·7°	+1·7°	+1·2°C	(120 cases)
	Siberia	+0·4	+0·3	+0·1	(103 cases)
w	Europe	−250	−350	−500 m/24^h	ca. 260 mb
	Siberia	−35	−45	−20	ca. 400 mb

If these somewhat dubious results are correct, we may conclude that cold-surface anticyclones, overrun by a cyclonic wind field above 2 km, possess only a small vertical and ageostrophic circulation, and that Fausts "zero-level"—in which maximum horizontal velocity of wind coincides with a change in sign of the vertical component ($w = 0$)—is situated lower than in middle latitudes, together with the low polar tropopause below 300 mb.

REFERENCES

[1] FLOHN, H. and TRENKLE, H. *Arch. Meteor. Geophys. Bioklim.* Vol. A7, pp. 85–98, 1954

[2] DEFANT, F. *Arch. Meteor. Geophys. Bioklim.* Vol. A6, pp. 253–279, 280–296, 1954

[3] SIMILÄ, A. *Vortrag Tagung Dt. Meteor. Ges. Berlin*, 28 September to 4 October 1953

[4] FLOHN, H. *Met. Rund.* Vol. 5, pp. 81–87, 121–128, 1952

[5] BREZOWSKY, H. *Ber. Dt. Wetterdienst* US-Zone Vol. 35, pp. 48–56, 1952

[6] FAUST, H. *Ber. Dt. Wetterdienst* US-Zone Vol. 35, 92–99, 1952; *Met. Rdsch.* Vol. 6, pp. 6–15, 1953

[7] SEIDEL, G. *Ber. Dt. Wetterdienst* Vol. 34, 1957

WINDS AND TEMPERATURES IN THE ARCTIC STRATOSPHERE

H. WEXLER and W. B. MORELAND
U.S. Weather Bureau, Washington, D.C.

More abundant aerological data require considerable modification of the early view of the temperatures and winds in the Arctic stratosphere. The availability of several months of daily charts at 50 mb (20 km) reveal cold cyclones and warm anticyclones which have little or no connection with pressure systems at lower levels. Vertical motions seem to be dominant in determining the temperature contrast between these 50 mb cyclones and anticyclones which may amount to 20° or 30°C. These slowly moving large cells tend to shift their prevailing locations from one year to the next, thus explaining why some investigators have concluded that the winter Arctic stratosphere is warm and others that it is cold.

Soundings made in the cold south-west quadrants of the cyclones at 50 mb show winter features similar to those in the Antarctic stratosphere: lack of a well-defined tropopause; slow decrease of temperature with height above about 10 km, reaching values of −75°C and lower at the top of the sounding (16–18 km). Despite these similarities, however, the Arctic stratosphere, represented by individual extreme soundings, appears to be 5–10°C warmer than monthly mean soundings at equivalent latitudes in the Antarctic (71–78°).

In summer the Arctic and Antarctic stratospheric temperatures are more nearly equal, with the possibility of the Antarctic being slightly warmer.

There is some evidence of a strong winter jet-stream (the Polar-Night Jet) which exists near 30 km at the boundary between the dark and sunlight stratosphere. Whether this is a property of the Sub-Arctic as a whole or whether it is found at the southern peripheries of the cold cyclones observed at 20 km is a matter for further investigation. If there does exist at 20–30 km a strong zonal westerly jet, then the extraordinarily large increase in total ozone content observed in late winter at Tromsö (latitude 70°) may perhaps be explained by a breakdown of these zonal winds into a blocking pattern whose large meridional motions transport ozone from lower latitudes. The large blocking anticyclones frequently found near Scandinavia at that season may serve as "sinks" into which ozone advected from lower latitudes converges and subsides.

In summer the easterlies replace the westerlies at heights above 17 km, but it is not likely that these summer easterlies extend to the North Pole.

MORE abundant aerological data available from the Arctic in recent years have contributed greatly to knowledge of the temperatures and winds of the Arctic stratosphere. Many of the early views, based on extrapolation poleward and upward from the few shallow soundings available, require considerable modification, particularly in the winter season.

WINTER

Before 1933 most vertical cross-sections of the atmosphere along a meridian showed the temperature of the stratosphere increasing poleward. This reversal of the usual poleward temperature decrease found in the troposphere

was supposed to gradually cancel out, with increasing height, the polar cyclone, so that near 20 km isobaric surfaces were believed to be essentially horizontal.[5]

In 1933, however, a vertical north-south cross-section published in *Physikalische Hydrodynamik*[3] showed at higher latitudes a poleward decrease of the stratospheric temperature.

Then in 1934, PALMÉN,[17] studying soundings made at Abisko, Sweden, at latitude 68° 21′N, showed that the average January temperature decreased from −55° at 55° latitude to −65°C farther north. Palmén's picture was verified over a broader meridional sector by summarization of all available pre-World War II aerological data.[16] This January vertical cross-section showed a "warm island" of stratospheric air between latitudes 50° and 70°, with a weak indication of decreasing temperatures to the north.

The most recent compilation of a January vertical cross-section (*Fig. 1*), based on observations made on the 80th West meridian during 1948–51, shows again the "warm island" of −50°C air centered near latitude 60° with temperatures at 16 km decreasing to −70°C over the Pole and at 30 km to an extrapolated −85°C.[12]

While at first sight Kochanski's 80°W vertical cross-section lends further substantiation to the cold Arctic stratosphere shown in Palmén's 20°E cross-section, the former cross-section has temperatures up to 20° colder above 20 km in the neighborhood of the Pole. Admittedly, a comparison is not quite fair since the 80°W section was based on data available much farther north (to 82°N) than the 20°E section (to 68°N), but it would be interesting to see if there could possibly be a true climatological temperature difference between these two meridians at 20 km in the Arctic stratosphere.

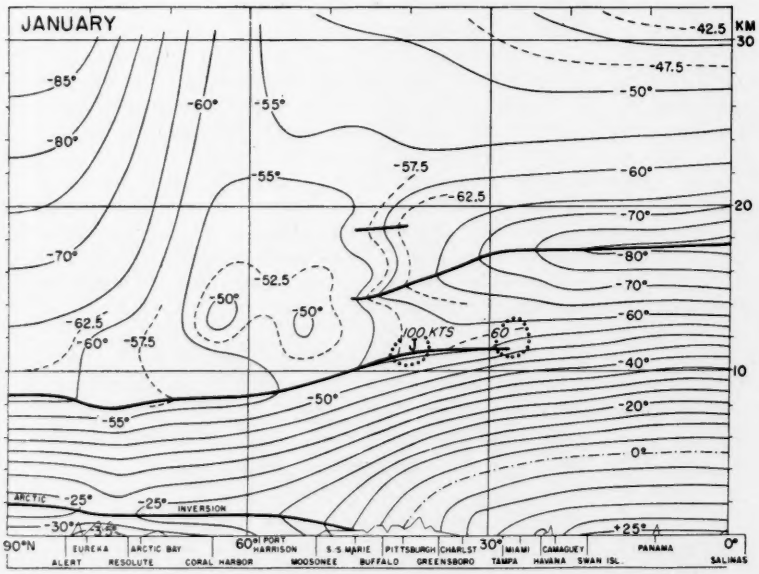


Fig. 1. Vertical cross-section along 80° W, January. Isotherms in °C (after KOCHANSKI[12])

Suppl.
Part.I
1956

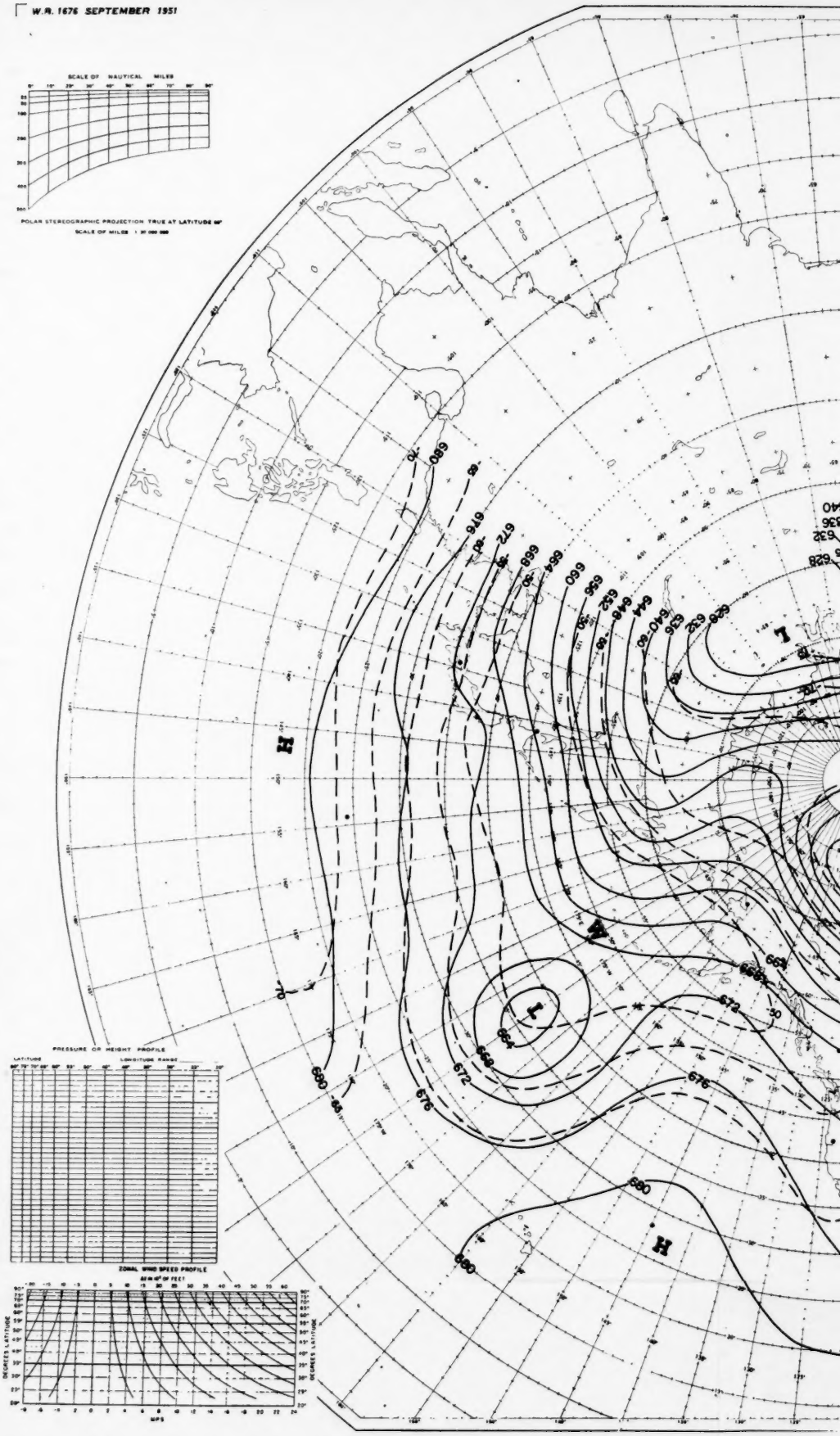
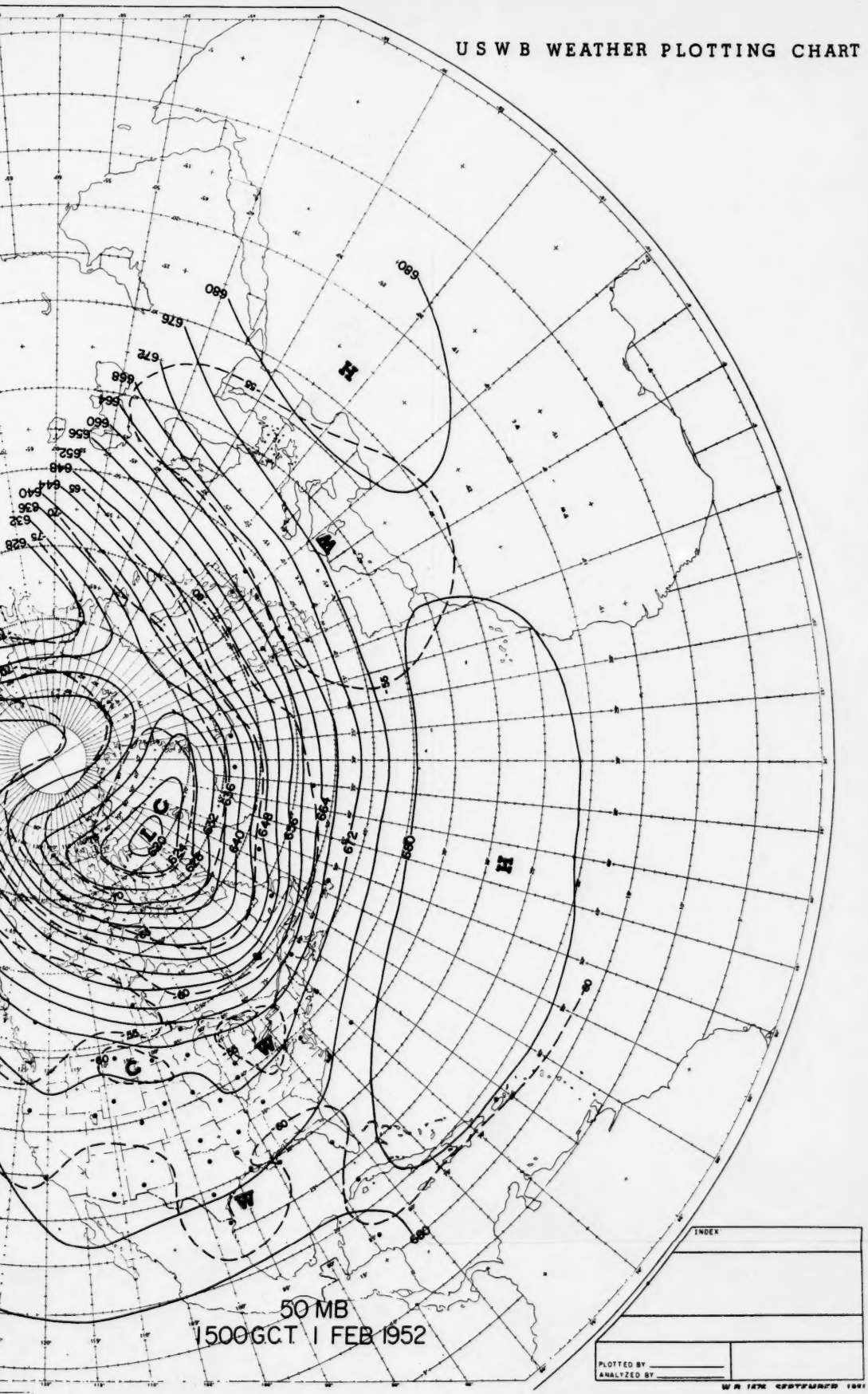


Fig. 2. 50-mb chart, 1 February 1952. Isothe



Isotherms in °C. Contours labeled in hundreds of feet

(*Facing page* 72)

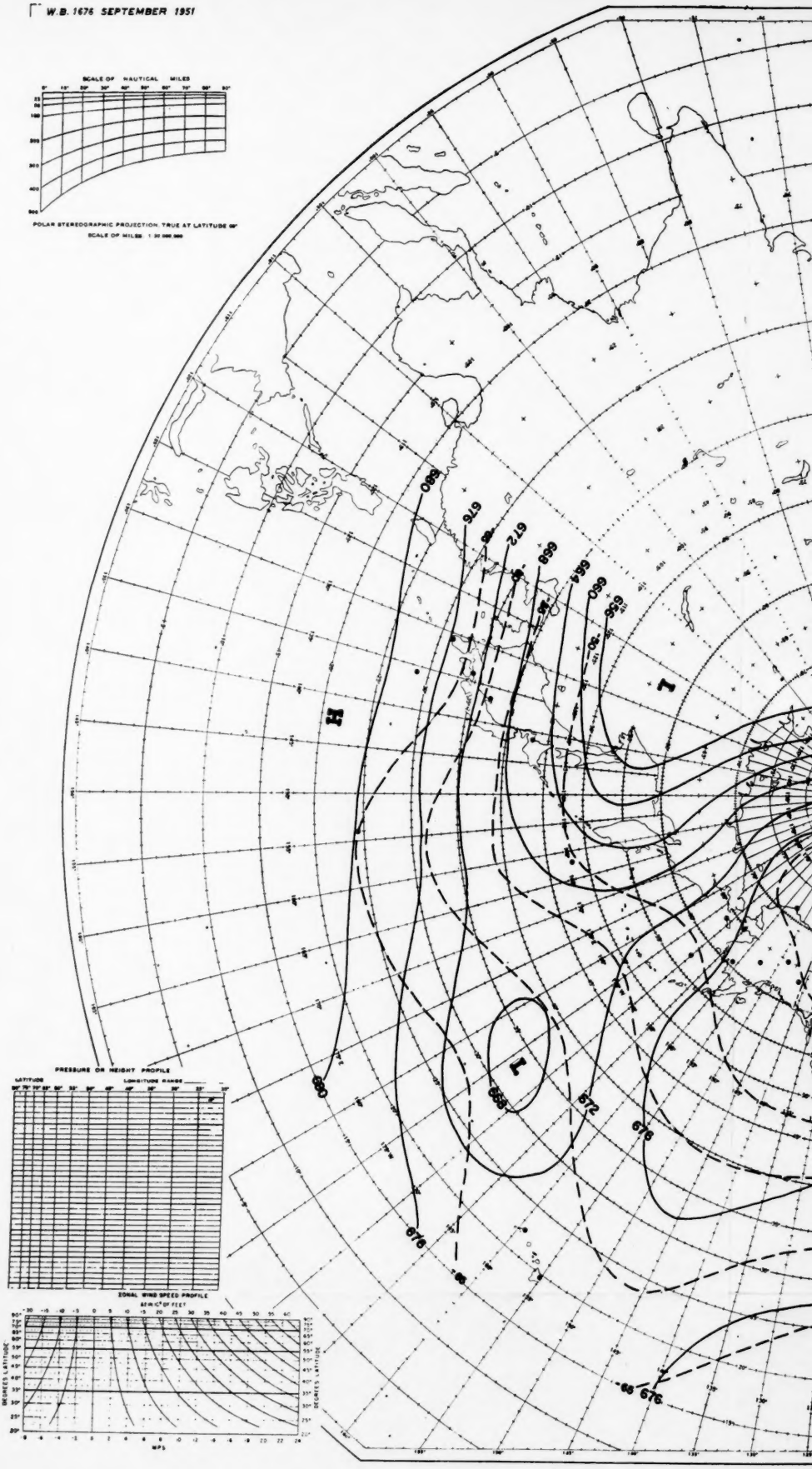
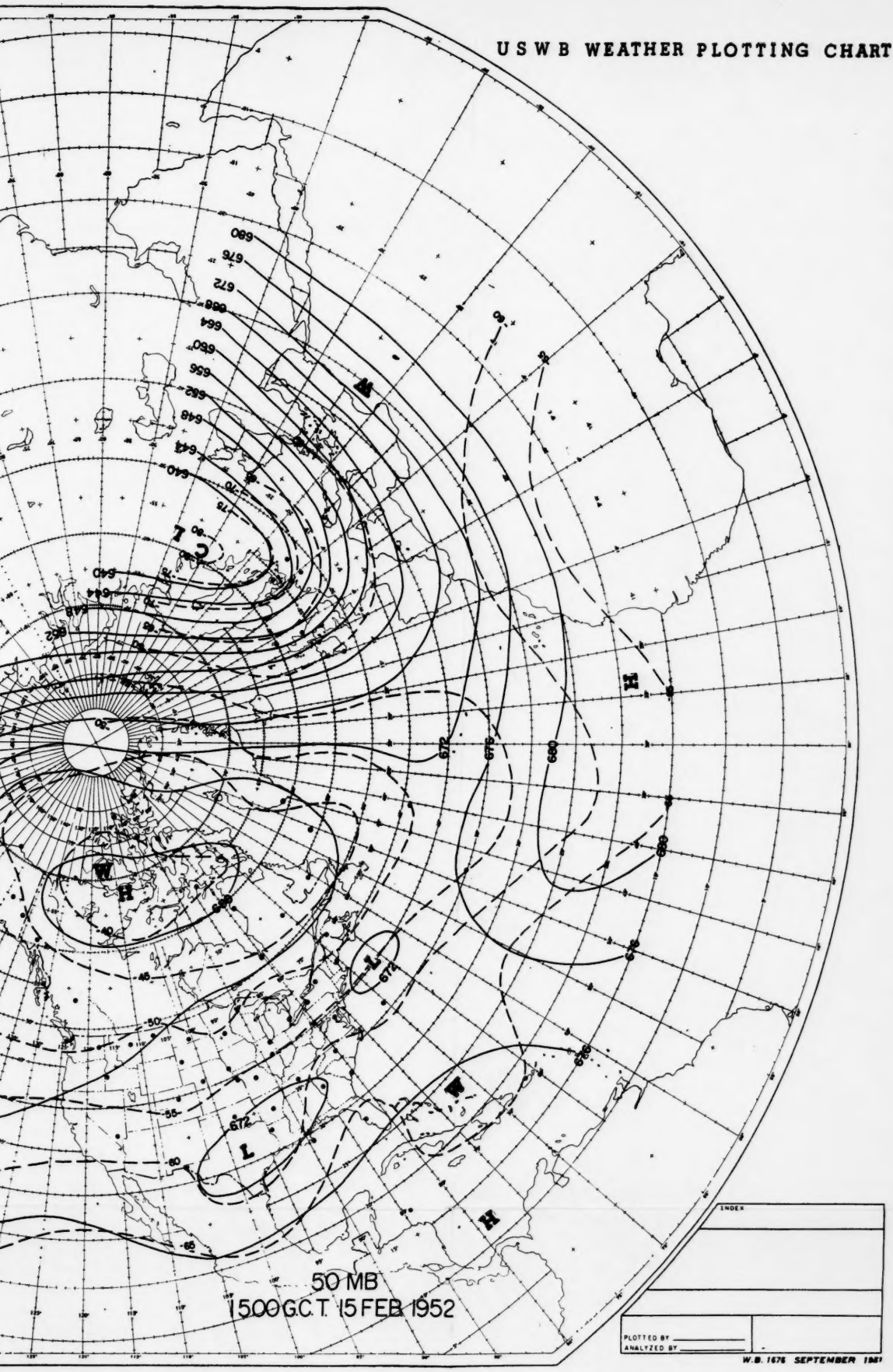


Fig. 3. 50-mb chart, 15 Fe



15 February 1952. Isopleths same as Fig. 2

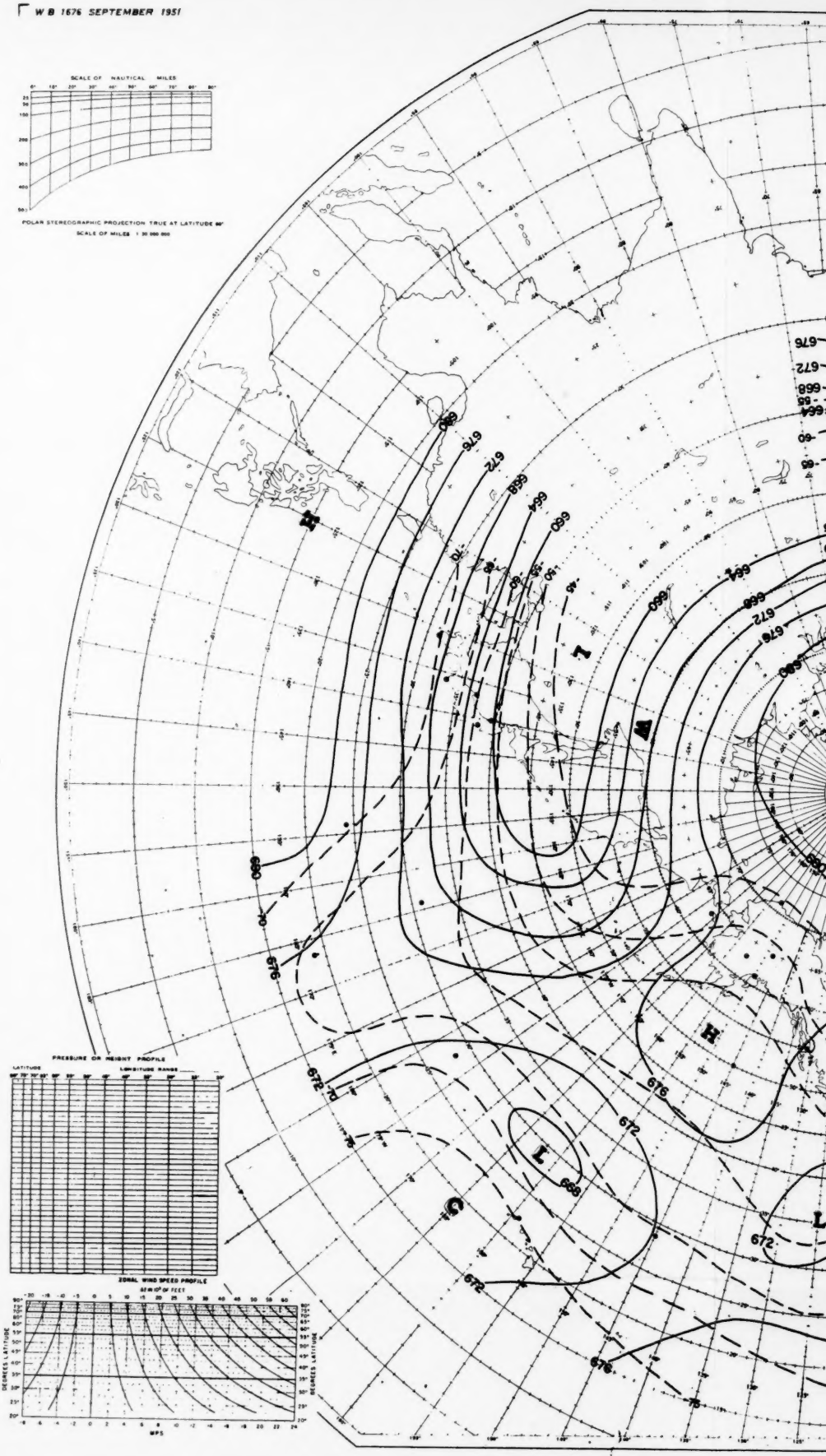
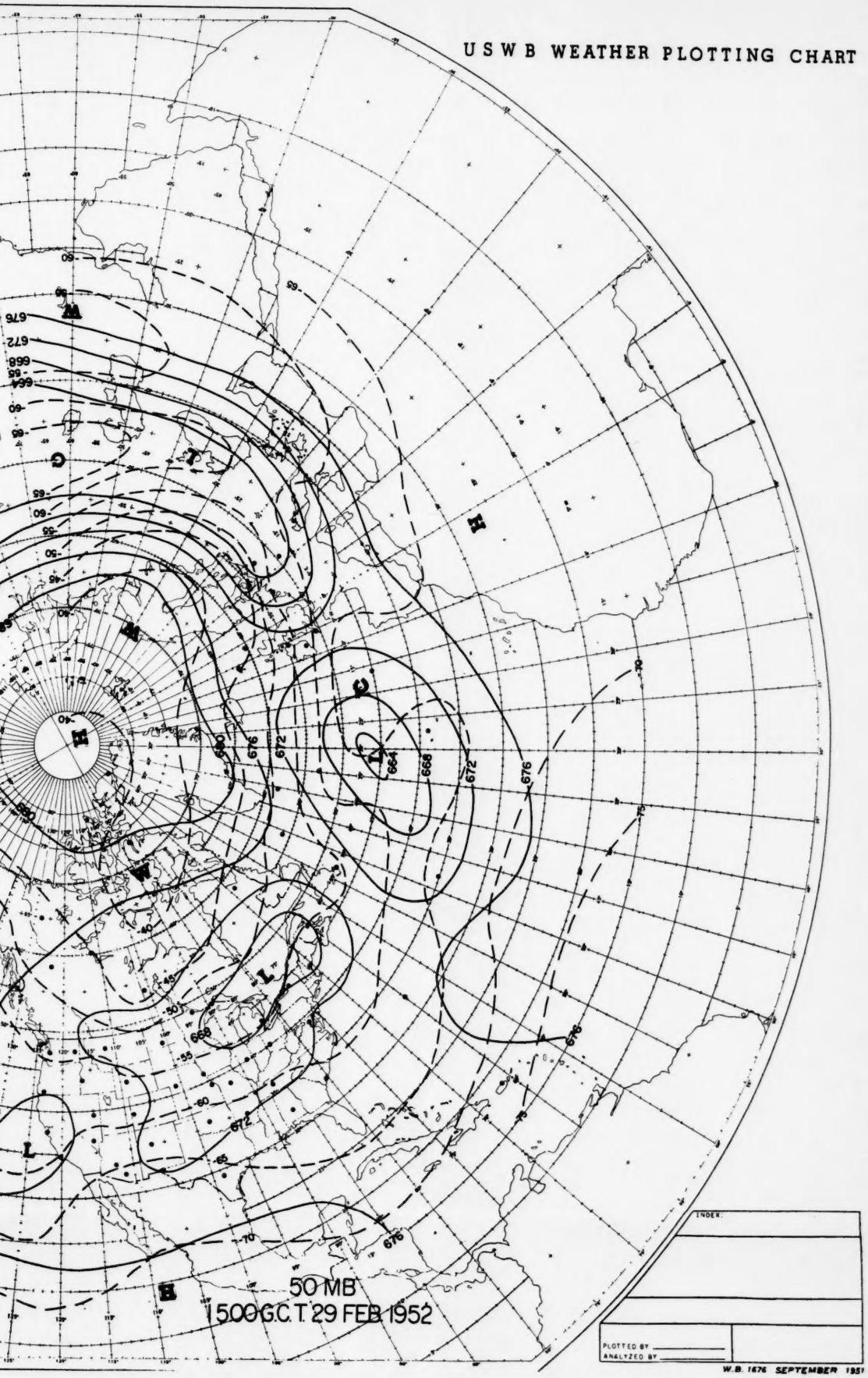


Fig. 4. 50-mb chart, 29 Febr



February 1952. Isopleths same as Fig. 2

Sup
Par
195

As a first step in the examination of this question, let us look at three nearly circumpolar charts of the 50-mb surface (~20 km) for 1, 15 and 29 February 1952. These charts are part of a series of daily charts from 26 January to 29 February 1952, drawn by Mr. F. Pooler, Jr., of the U.S. Weather Bureau, to look into the synoptic background of the famous Berlin "warming" of 23 February 1952. These charts, discussed in greater detail elsewhere,[22] show the existence of considerable activity at a level once thought to be essentially motionless.

The first chart (*Fig. 2*), 1 February, shows a strong cyclone centered over Baffin Bay with temperatures below −75°C and a weak anticyclonic ridge located over the Gulf of Alaska, temperatures near −50°C. In two weeks, as illustrated by the 15 February chart (*Fig. 3*), the Baffin Bay cyclone, still with −70°C or lower temperatures, moved south-eastward over northern Europe, while the anticyclonic ridge moved eastward into central Canada and developed into a large anticyclonic eddy, warming by 10°–15°C, presumably by subsidence. By 29 February (*Fig. 4*) the cold cyclone moved into south-east Europe closely followed by the warm anticyclone. The strong horizontal gradient of temperature thus created moved over Berlin and was probably responsible for the rapid Berlin warming.

A second series of daily Northern Hemisphere synoptic charts at 50 mb, drawn for seven months of the year 1953, was published in 1955 as part of the Weather Bureau historical map series, with support by the Geophysics Research Directorate of the U.S. Air Force, Cambridge Research Center.[15] Techniques of data checking, extrapolation and analysis are given in a paper by MORELAND and CLUFF.[14]

A portion of the circumpolar 50-mb chart for 15 January 1953 is shown in *Fig. 5* above the January mean (1948–51) chart prepared by KOCHANSKI.[11] There is very little resemblance between the two charts, but there is a quite close similarity between Kochanski's mean January chart and that of 1 February 1952 (*Fig. 2*), which is fairly typical of the charts during the latter part of January 1952 (not enough charts from January were available to construct a mean January 1952 chart). The inference is that the January 1952 50-mb pattern was probably quite close to the 1948–51 average, but that the January 1953 pattern was not. If the "normal" position of the deep cyclone at 50 mb is near north-west Greenland, as determined by the 1948–52 period, then January 1953 was decidedly abnormal, the cyclone being located near northern Scandinavia.

In the winter of 1953–54, judging from the several series of three-day mean charts at 50 mb drawn by AUSTIN and KRAWITZ,[1] the cyclone was back in its "normal" position near north-western Greenland.

Returning to the daily 50-mb Northern Hemisphere maps for 1953, which is the longest known series of such charts, further analysis confirms the outstanding synoptic features at this level first revealed by the 1952 charts, namely, that at higher latitudes there exist a few large cyclones and anticyclones which move very slowly compared to those at lower levels. These two factors account for far fewer 50-mb pressure centers crossing selected meridians, as illustrated in *Table 1.*

This striking difference in the number of passages is not caused by lack of data at the top level, since there was fairly good data coverage at the three

Suppl. Part. I 1956

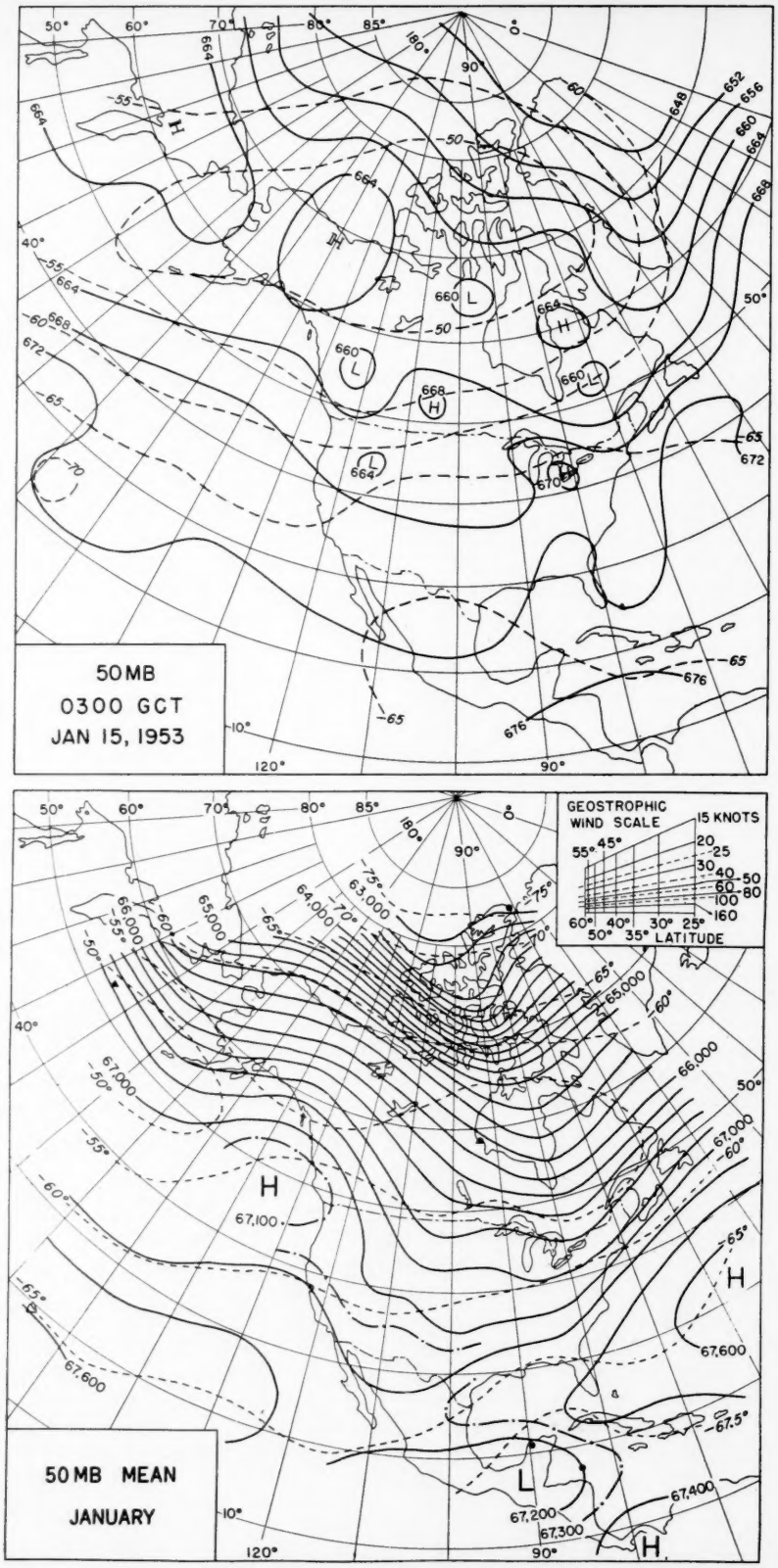


Fig. 5. A comparison of a daily 50-mb chart, 15 January 1953, with the 50-mb mean January chart prepared by KOCHANSKI.[11] *Contour interval 200 ft on mean chart, 400 ft on daily chart*

Table 1. Number of High and Low Centers Crossing the 180°, 90° W and 0° Meridians in a Latitude Belt from 40° N to the Pole, at the Surface, 500 mb, and 50 mb for January–March 1953

	180°		90° W		0°	
	High	*Low*	*High*	*Low*	*High*	*Low*
50 mb	3	9	1	14	1	10
500 mb	9	36	5	47	14	49
Surface	16	42	29	45	17	41

meridians selected. The 50-mb maps also showed quite strikingly that the south-west quadrants of these large, intense cyclones are quite cold (−70° to −80°C), much colder than the air farther north over the Arctic Basin. Also, the isotherms in that quadrant do not move with the wind from day to day, indicating that vertical motions exist in this south-west quadrant of the order of 1 cm/sec—first upward as the air moves from the north-west towards the colder isotherms, then downward. On the other hand, the anticyclones are much warmer, with temperatures of −40° to −55°C, presumably caused by subsidence.

It appears that the temperature distributions within the large cyclones and anticyclones are controlled largely by vertical motions which extend through a considerable thickness of the ozonosphere. If this is so, an independent check could be made by means of ozone measurements, more ozone to be noted in that portion of the south-west quadrant of a 50-mb cyclone where the air is descending and less ozone where the air is ascending.[9] Also, the warming anticyclonic cells should have higher ozone contents, such as was noted in mid-February 1952 over Edmonton, Canada.[22] With the increasing availability of 50-mb charts, it is hoped that this hypothesis can be given adequate testing soon on the relatively dense western European network of ozone observing stations.

Returning to Kochanski's January mean 50-mb chart (*Fig. 5*), we see a zone of strong winds in northern Canada (maximum 110 knots at 75°N, 120°W), which is also borne out by the 1 February 1952 chart (*Fig. 2*). This wind maximum, which has been called the "polar-night jet", was apparently first discovered by FLOHN[7] in constructing his pole-to-pole vertical cross-sections for winter and summer. In both winter hemispheres the polar-night jet was found by Flohn to be near latitude 60° at 30 km, reaching average speeds of 60 knots, or as strong as the more familiar low-level jet found near 30° latitude at 10 km.

We believe that these polar-night jets are *not* circumpolar, and that they are probably found only in the southern portions of the deep cyclones at 20 km which undoubtedly extend also to the 30-km level. The location of these jets will therefore change with the position of the deep cyclones to which they are attached, as illustrated by the daily charts in *Fig. 2* (mid-winter 1952) and *Fig. 5* (mid-winter 1953).

There is another, more indirect, type of evidence which may indicate a shift in these upper-level patterns from one winter to another, namely, the seasonal variation in the total ozone content of the vertical air column, most of which is found in the layer from 15 to 25 km high. In *Fig. 6* are shown 4 smoothed curves illustrating the seasonal variation in ozone at 4

Suppl. Part. I 1956

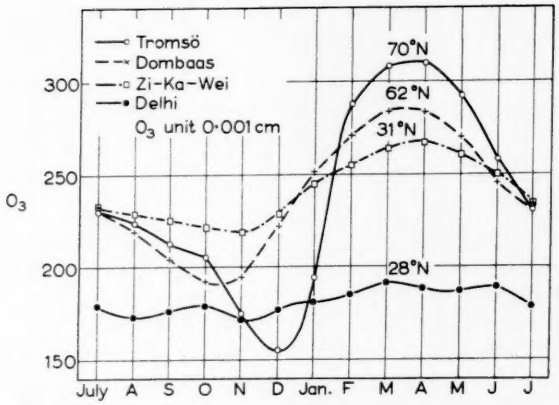


Fig. 6. Seasonal variation in ozone content at 4 latitudes (after LANGLO[13]*)*

different latitudes ranging from 28° to 70°N.[13] Note the large increase in ozone at Tromsö (latitude 70°) from the lowest value of all latitudes in December to the highest value in March and April. Although there may be some doubt as to the actual value of the low Tromsö ozone content in the dark months of November, December and January, there appears no reason to doubt the high values observed in February, March and April, when more accurate ozone measurements can be made in sunlight.

Our hypothesis for this extraordinary rise in ozone at Tromsö is the following: As summer is succeeded by autumn, a net radiative cooling in the stratosphere occurs at high latitudes, thus changing the summer stratospheric easterlies to westerlies at middle and high latitudes. This process is illustrated in *Fig. 7*, the changes in monthly profile of pressure at 19 km.[21] In the dark portion of the stratosphere no new ozone is being manufactured, and if the flow is essentially zonal, then little ozone is being transported from lower

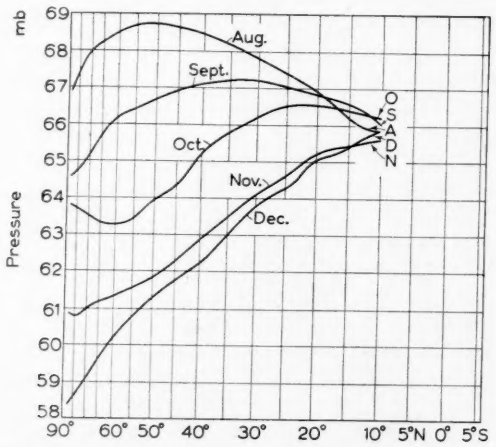


Fig. 7. Normal monthly pressure profiles at 19 km. Abscissa is sine of the latitude (after WEXLER[21]*)*

latitudes (*Fig. 8*). Meanwhile, the ozone which was originally in the air gradually sinks with the cooled air to lower levels and spasmodically leaks off southward in surface flows of polar air, thus causing a pronounced ozone minimum at high latitudes.

However, as the cooling proceeds, the meridional temperature gradient and the stratospheric west winds become stronger until a large-scale instability occurs, causing formation of a deep cyclone near north-western Greenland and an anticyclone farther east over Scandinavia—a process reminiscent of the formation of "blocking patterns" found in the troposphere.[2] This breakdown of essentially zonal flow into large-scale meridional flow encourages the transport of ozone from low to high latitudes into the stratospheric anticyclones which then become natural "traps" for the ozone feeding into them from lower latitudes. The entrapped ozone gradually sinks

Suppl.
Part.I
1956

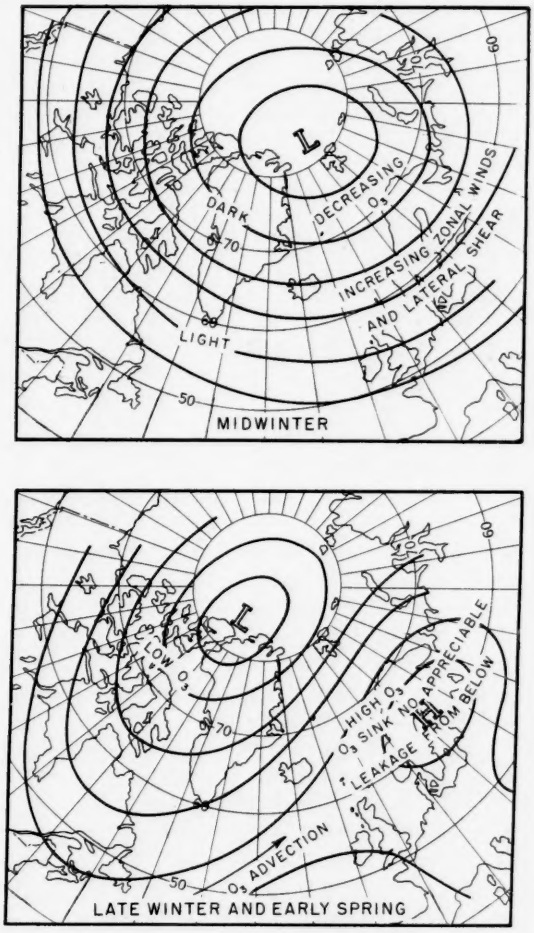


Fig. 8. Sketch illustrating the change in the flow at the 50-mb level from mid-winter to late winter and early spring and its effect on the ozone content

to lower altitudes but does not leak off in any significant amount from the closed dynamic anticyclonic cell. This accumulation of ozone from lower latitudes, plus that created photochemically by ultraviolet radiation from the returning sun, accounts for the late winter and early spring Tromsö ozone amounts being larger than those at lower latitudes, despite the longer periods of solar radiation at lower latitudes.

If this explanation of the Tromsö late winter ozone rise to the highest value shown in *Fig. 6* is correct, then it follows that this rise need not be found everywhere along the 70° latitude circle. Only those regions located under the "blocking" stratospheric anticyclones should have high ozone contents while those located under the stratospheric cyclones should have low values. A test of this hypothesis will occur during the International Geophysical Year from ozone measuring instruments which have been installed at Alert (82° 30′N, 62°W) and Resolute Bay (75°N, 95°W), in the vicinity of the "normal" position of the 50-mb cyclone.

Returning to the sharp winter increase at Tromsö, Langlo has shown (*Fig. 9*) that in 1948 this occurred about one month later than in 1940.[13] Unfortunately, there do not exist sufficient high-altitude aerological data

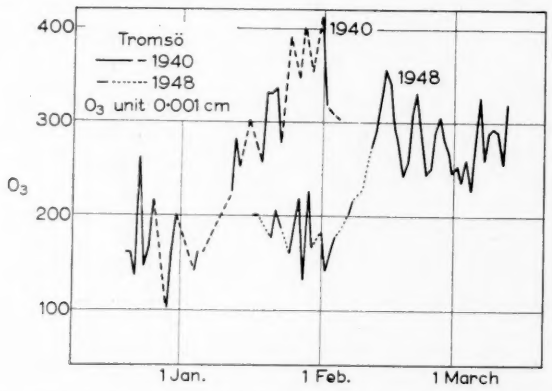


Fig. 9. Variation in time of winter increase of ozone at Tromsö (after LANGLO[13])

to see if there were significant differences in 50-mb flow patterns between those 2 years. But again, reasoning from the hypothesis outlined earlier, one would suspect that since the 1940 major ozone rise occurred in the latter half of January, this was a normal year in so far as the 50-mb flow pattern was concerned, with the cold low probably located over north-west Greenland. The year 1948 with its late ozone rise was probably an "abnormal" year similar to that of 1953 when the cold low was centered north of Scandinavia through the month of January and part of February. It returned to its "normal" position in the latter part of February with a ridge forming over western Europe.

A check of the mean monthly total ozone values for Aarhus, Denmark (56°N, 10°E), for the months October–May, 1951–52 and 1952–53 (*Fig. 10*), also indicates that the 1952–53 winter was probably anomalous. The curve for 1951–52 shows the usual seasonal rise of ozone values, as indicated by

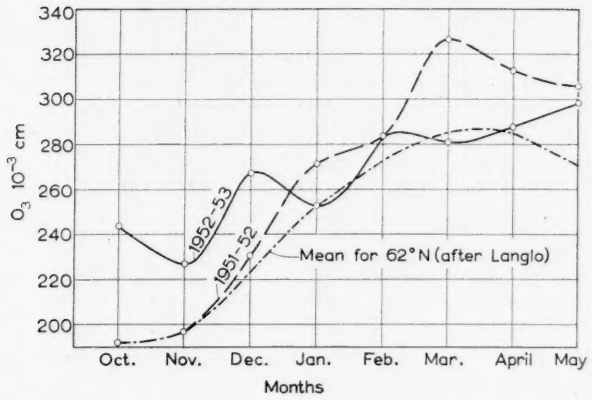


Fig. 10. Variation in the seasonal rise of ozone for the years 1951–52 and 1952–53 at Aarhus, Denmark

the mean curve for 62°N,[13] while the curve for 1952–53 shows the seasonal rise being retarded from December to January, and then rising again in late winter and early spring.

Before leaving the winter aspects of the Arctic stratosphere, let us look at some temperature-height curves from the Arctic and Antarctic[23] shown in *Fig. 11.* The Little America and Maudheim curves show the typical absence of a well-defined tropopause which was first noted by COURT.[4] The Arctic

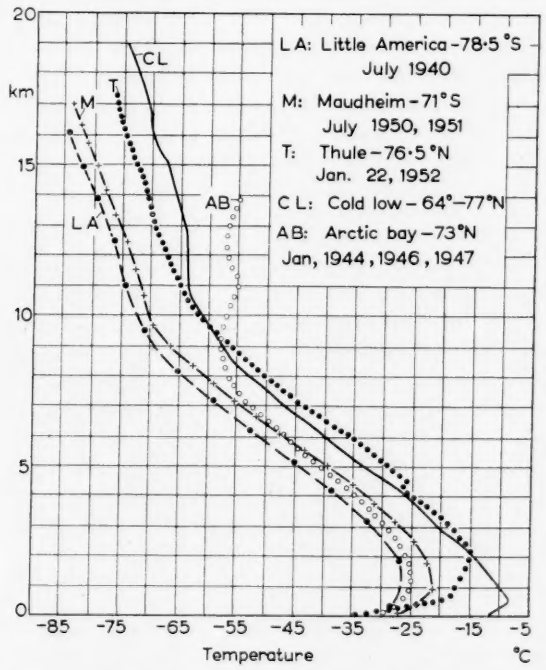


Fig. 11. Winter soundings, Antarctica and the Arctic (after WEXLER *and* RUBIN[23])

Bay (73°N, 85°W) sounding, formerly thought to be typical of the Arctic, has a troposphere of about the same temperature as Little America and Maudheim, but a much warmer stratosphere (−55°C vs. −75°C). No doubt some of this difference is due to the fact that in the stratosphere the Arctic Bay sounding is probably biased toward the warmer temperatures. It has been our experience that in the region of a stratospheric cold low, very few soundings reach even the 100-mb level. Thus one must be careful when making comparisons of different soundings which are based only on a limited number of observations.

Now, recalling the previous discussion concerning the 50-mb cyclones and anticyclones, we find their pronounced temperature differences render uncertain what is meant by a "typical" Arctic stratosphere. If an average sounding is composed of those made in the south-west quadrants of 50-mb cyclones, then the curve "CL" results, which also has an ill-defined tropopause and is 10°C colder than the Arctic Bay stratosphere. The Thule sounding of 22 January 1952, taken near the cyclonic center, is also quite parallel to but warmer than the Antarctic soundings.

SCHUMACHER[19] has attempted to explain the colder Antarctic winter stratosphere by the much colder surface temperatures (some 30°C lower than those in the Arctic); this appreciably reduces the intensity of the 9·6μ infrared radiation which is absorbed by the ozone in the stratosphere. However, it would be difficult to explain by this theory the very cold stratospheric temperatures found at Thule, Greenland, in late January 1952 when values as low as −81°C were observed with surface temperatures no lower than −28° to −38°C within hundreds of miles radius. Likewise, when surface temperatures in north-eastern Siberia fall to values comparable to those observed in Antarctica, the stratospheric temperatures are much higher than those found over Maudheim in Antarctica. For example, FLOHN[8] shows that at Yakutsk (62°N, 130°E) during the winter of 1950–51, the surface temperature averaged −43°C but the 200-mb temperature was only −59°C, or 13° warmer than the Maudheim winter temperature at the same level.

It would appear that vertical motions associated with major features of the flow patterns cannot be ignored in explaining the temperature distribution within the two polar stratospheres. It may be that the very low temperatures observed in the Little America and Maudheim winter stratospheres are associated with cold stratospheric cyclones in those areas and that other portions of the Antarctic stratosphere which are dominated by intervening anticyclones have higher temperatures. The radiosonde observations at Port Martin (67°S, 141°E) show higher stratospheric temperatures than Little America (78°S, 164°W) and Maudheim (71°S, 11°W), but some of this may be a latitude effect.[23]

However, even if it should turn out that Little America and Maudheim are, on the average, under cold winter stratospheric cyclones, the fact that their average stratospheric temperatures are colder by some 10°C than those observed in the Arctic cold cyclones (e.g. Thule) may point to the importance of a surface radiation effect as suggested by Schumacher. To really confirm this, one must have sufficient high-altitude wind data to construct trajectories of stratospheric air. Both Maudheim and Little America are so close to the coast that on certain days the stratospheric air may have

travelled long distances over the open ocean whose more intense infrared radiation should result in higher stratospheric temperatures if Schumacher's explanation is correct. (Schumacher's sounding IV, in his *Fig. 1*, made on 13 August 1950, has a high temperature of −8°C at 850 mb, indicating a recent maritime source, at least of the lower troposphere, yet the stratospheric temperature reaches a value of −91°C, the lowest observed at Maudheim.)

Summer

In summer the poleward increase of stratospheric temperatures has been known for some time. Palmén argued from consideration of the Abisko soundings that the polar stratospheric temperatures were higher than formerly thought (temperatures as high as −35°C at 20 km), but later Johansen,[10] referring to the Tromsö soundings, confirmed Väisälä's findings that because of radiation error the Abisko soundings were too warm by some 5°C at the top levels of the summer soundings; that is, the temperatures at 13–20 km are about −40° in June and July in northern Scandinavia (~70° latitude). The July hemispheric cross-section prepared by Namias and Gentry[16] shows polar stratospheric temperatures of about −45°C, and Kochanski's 80°W section has temperatures somewhere between −40° and −45°C.

Suppl.
Part. I
1956

In *Fig. 12* there is shown a series of summer soundings for the Arctic and Antarctic taken from Flohn.[6] Here the Antarctic soundings (Little America and Schwabenland) have colder tropospheres than their Arctic counterparts (Tromsö, Point Barrow and Spitsbergen), but slightly warmer stratospheres (even allowing for possible radiation errors in the early radiosondes used at Little America and Schwabenland). Sverdrup[20] suggests that the higher surface albedo in Antarctica accounts for the colder summer troposphere there as compared to the Arctic. It is likely that this same higher albedo also accounts for the slightly higher temperatures of the Antarctic stratosphere in the following way: The high intensity of the reflected solar radiation can warm the stratosphere by absorption of the Chappuis bands by the ozone. Rough calculations based on a surface albedo difference of 80 per cent shows a higher temperature of the stratosphere over the higher surface albedo by $\Delta T = 1{\cdot}5/K$ °C, where K is the emissivity of the absorbing layer. If K is 0·5, as appears reasonable, then $\Delta T = 3$°C which is in good agreement with the soundings shown in *Fig. 12*. The albedos in the neighbourhood of the Arctic stations used in that figure, located as they are near bare land and open ocean in summer, should be much lower than the Little America and Schwabenland soundings taken above or near extensive snow and ice fields of Antarctica.

During the summer months the circulation in the Arctic stratosphere between 17 and 21 km is quite weak. Geostrophic winds obtained from daily 100-mb and 50-mb charts for July 1953 indicate that speeds average from 5 to 15 knots. Wind directions to a large extent are variable.

Mean meridional cross-sections published by Flohn[7] and by Kochanski[12] for July and summer, respectively, have shown a circumpolar anticyclone above 14–16 km, with easterly winds extending southward to the equator. Also the July mean 41-mb chart, prepared by Scherhag,[18] indicates

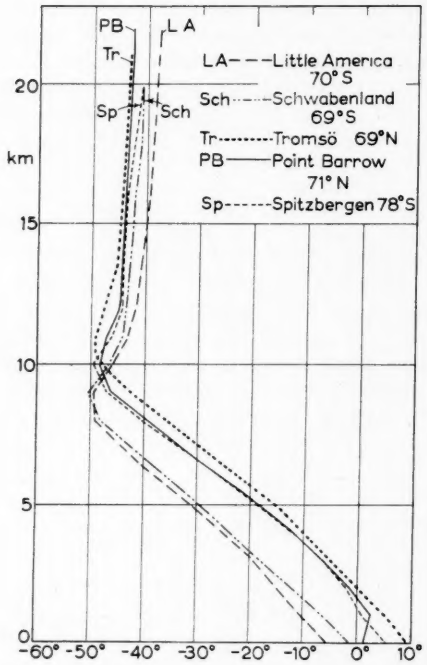


Fig. 12. Summer soundings, Antarctica and the Arctic (after FLOHN[6])

easterlies in the vicinity of the North Pole. However, more recent data give evidence that, in the central Arctic region during the summer, westerly winds may exist at heights above 20 km. The pressure profile at 19 km for the month of August (*Fig. 7*) shows a decrease of pressure from 60°N to the Pole and thus a westerly circulation in this circumpolar belt. Further corroboration of these westerlies is found on the daily July 50-mb charts for 1953.[15] Although a mean chart for this month has not been constructed, an inspection of the daily maps points out the absence of a circumpolar high. Instead, a belt of high pressure, consisting of a number of individual pressure centers, is found around the hemisphere between 70° and 75°N. Thus this series of summer daily charts at 50 mb strongly suggests that in summer a narrow belt of westerlies exists between the Pole and 80°N up to heights of 21 km. This belt of westerlies, however, weakens with height. Thermal wind considerations indicate that near 30 km the winds become easterly around the Pole.

REFERENCES

[1] AUSTIN, J. M. and KRAWITZ, L. "50 Millibar Patterns and their Relationship to Tropospheric Changes" *J. Met.* Vol. 13, No. 2, pp. 152–159, 1956

[2] BERGGREN, R., BOLIN, B. and ROSSBY, C.-G. "An Aerological Study of Zonal Motion, its Perturbations and Break-down" *Tellus* Vol. 1, No. 2, pp. 14–37, 1949

[3] BJERKNES, V. and J., SOLBERG, H., BERGERON, T. *Physikalische Hydrodynamik;* J. Springer, Berlin, 1933

[4] COURT, A. "Tropopause Disappearance during the Antarctic Winter" *Bull. Am. met. Soc.* Vol. 23, No. 5, pp. 220–238, 1942

[5] EXNER, F. *Dynamische Meteorologie*, 2nd Ed., p. 43; Julius Springer, Wien, 1925

[6] FLOHN, H. "Grundzuge der Allgemeinen Atmosphärischen Zirkulation auf der Südhalbkugel" *Archiv. f. Met., Geophys., u. Bioklim.* Ser. A, Bd. II, H.1, pp. 17–64, 1950

[7] ——. "Die Planetarische Zirkulation der Atmosphäre bis 30 km Höhe" *Ber. Deutsch. Wetterdienst.*, No. 12, pp. 156–161, 1950

[8] ——. "Zur Aerologie der Polargebiete" *Met. Rund.* Vol. 5, No. 6, pp. 81–87, 1952

[9] GÖTZ, F. W. P. "Ozone in the Atmosphere" *Compendium of Meteorology;* American Meteorological Society, Boston, Mass., 1951

[10] JOHANSEN, H. "Mean Upper Air Data, Tromsö, 1941–44" *Geofys. Publ.* Vol. XVII, No. 4, 1949

[11] [KOCHANSKI, A., and WASKO, P.] "Mean Monthly Maps of 300, 200, 100, 50 and 25-mb Surfaces over North America" *Air Weather Service Technical Report* No. 105–107, Washington D.C., 1953

[12] KOCHANSKI, A. "Cross-sections of the Mean Zonal Flow and Temperature along 80°W" *J. Met.* Vol. 12, No. 2, pp. 45–106, 1955

[13] LANGLO, K. "On the Amount of Atmospheric Ozone and its Relation to Meteorological Conditions" *Geofys. Publ.* Vol. XVIII, No. 6, 1952

[14] MORELAND, W. B., and CLUFF, F. D. "Problems and Examples of Northern Hemisphere Stratospheric Analyses" Paper presented at the 136th National Meeting of the American Meteorological Society; Washington, D.C., 5 May 1955

[15] [MORELAND, W. B.] *Daily Series Synoptic Weather Maps, Northern Hemisphere, 100-mb and 50-mb Charts;* U.S. Weather Bureau, Washington, D.C.,1955

[16] [NAMIAS, J., and GENTRY, R. C.] *Normal Weather Maps, Northern Hemisphere Upper Level, 10,000 ft, 20,000 ft, 10, 13, 16 and 19km;* U.S. Weather Bureau, Washington, D.C., 1944

[17] PALMÉN, E. "Über die Temperaturverteilung in der Stratosphäre, etc." *Met. Z.* Bd. 51, pp. 17–23, 1934

[18] SCHERHAG, R. *Neue Methoden der Wetteranalyse*, pp. 424; Springer-Verlag, Berlin, 1948

[19] SCHUMACHER, N. J. "Upper Air Temperatures over an Antarctic Station" *Tellus* Vol. VII, No. 1, pp. 87–95, 1955

Suppl.
Part.I
1956

[20] SVERDRUP, H. U. "Some Problems in Arctic Meteorology" *Proc. Toronto Met. Conf. 1953;* published by R. Met. Soc., pp. 69–73, 1954

[21] WEXLER, H. "Spread of the Krakatoa Volcanic Dust Cloud as Related to the High-level Circulation" *Bull. Amer. met. Soc.* Vol. 32, pp. 48–51 1951

[22] ——. "A Meteorologist Looks at the Upper Atmosphere" Talk presented at the 250th Benjamin Franklin Anniversary, American Academy of Arts and Sciences, Boston, Mass., 11 January 1956. (To be published in Franklin Symposium volume *Atmospheric Explorations*, John Wiley, The Technology Press, M.I.T., 1958.)

[23] WEXLER, H., and RUBIN, M. "Antarctic Climatology and Meteorology" Talk presented at Symposium on the Antarctic, Amer. Geophys. Union, 26–27 April 1956, Washington, D.C. (Published in "Antarctica in the International Geophysical Year" *Geophysical Monograph* No. 1, Amer. Geophys. Union, pp. 36–43, 1956

Suppl.
Part. I
1956

CANADIAN STUDIES OF THE HIGH-LATITUDE STRATOSPHERIC JET STREAM IN WINTER*

WARREN L. GODSON
Meteorological Service of Canada

A research project has been organized in the Research Section, with R. Lee as project director, which has a two-fold purpose: (a) to establish the existence of the high-latitude stratospheric jet stream at and above the 18-km level as a synoptic entity; and (b) to study the hypothesis that its existence is controlled primarily by differential solar-radiational heating along the fringe of polar darkness near the level of maximum concentration of ozone. These problems are being attacked by means of synoptic cross-sections of the stratosphere in the Canadian Arctic, utilizing measured winds exclusively and by a systematic analysis of the winter 100-mb temperature field at 11 Arctic and Sub-Arctic stations situated between 64°N and 83°N.

Several selected vertical cross-sections are presented to demonstrate the existence of the jet stream as a synoptic entity. Hitherto, the existence of a mean jet stream above 18 km at high latitudes has been inferred indirectly from mean stratospheric temperatures and hydrostatic considerations rather than from direct wind observations. Cross-sections for November and early December 1955 show the jet stream at successively lower latitudes, consistent with the southward movement of the zone of polar darkness.

Evidence in support of the stated hypothesis (b) is presented from 10-day running mean temperatures for the 1955–56 winter at selected Arctic and Sub-Arctic stations. In October, running mean 100-mb temperatures over the latitude range 64–83°N all lay within a range of 2 Celsius degrees. The mean temperatures fell uniformly at all stations as the duration of daylight decreased. After the onset of polar darkness at Alert (83°N) at 20 km on 28 October, an increasing N–S temperature gradient developed due to differential heating along the fringe of polar night. Simultaneously, quasi-regular mean periodic temperature fluctuations appeared at all stations, suggesting the existence of waves in the baroclinic zone below the jet stream analogous to those found in the troposphere below the jet streams in middle latitudes. The amplitudes of these fluctuations generally increased with time until just before the vernal equinox, when they virtually disappeared except at Alert. After 7 March, when Alert received six hours of sunlight, temperatures at 100 mb throughout the Canadian Arctic increased from −65°C to around −45°C. Thereafter, a quasi-uniform temperature field of about −44°C was established at 100 mb in the latitude belt 64–83°N. Simultaneously, all signs of westerly winds as well as temperature fluctuations disappeared at these levels. Thus it can be inferred that differential solar heating of the ozone layer is primarily the physical process initiating the jet stream in the fall and that the reversal in the latitudinal input of solar radiation at the vernal equinox is effective in destroying it in early spring.

The stated hypothesis implies a mean southward movement of the jet stream until the winter solstice followed by a northward trend. There are four pieces of evidence in support of this statement: (1) direct evidence from 12 vertical cross-sections constructed

* This paper has been published in full in *Journal of Meteorology* Vol. 14, pp. 126–135, 1957.

for November and early December 1955; (2) the occurrence of two periods when maximum amplitudes in temperature fluctuations occurred at Alert; (3) the meridional shift in the maximum amplitudes of these temperature fluctuations, i.e. toward the south in December and January, while later they occurred at Alert when they disappeared at lower latitudes; (4) an analysis of the running variances of the 100-mb temperatures from their 20-day running means at Alert which showed two maxima, one in early December and one in mid-March.

Supp
Part
195

THE PROBLEM OF THE MAINTENANCE OF THE LOW-LEVEL ANTICYCLONIC CIRCULATION IN THE ARCTIC

RAGNAR FJØRTOFT
Det Norske Meteorologiske Institutt, Norway

The normal pressure maps for November are examined in light of the equations determining the distribution of vertical motions and the corresponding fields of horizontal divergence. The investigation suggests that the Arctic anticyclonic circulation in low levels is maintained essentially in the way that the flow patterns and mass distributions frequently existing in the atmosphere, and also found in the normal maps, cause a tendency for the air masses to be deflected in such a way to give a net warming by advection in the surroundings of the region where the Arctic anticyclone is found, thus causing the air in the middle to descend and spread at the ground.

Suppl.
Part. I
1956

1. INTRODUCTION

IT is well known that in Arctic and polar regions during the winter an anticyclonic circulation predominates in low levels. This anticyclonic circulation is, for instance, easily recognized on the normal surface pressure map for November (*Fig. 1*).

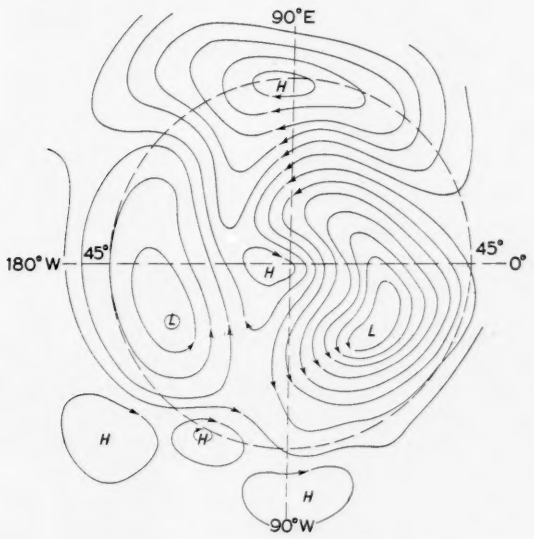


Fig. 1. Normal surface pressure map for November

When it comes to the problem of how this low-level anticyclonic circulation is maintained against the dissipating forces of friction it is generally referred to the effects of radiative cooling. In this note it will be shown (1) that the normal flow and temperature patterns, confining the investigation to the month of November, indicate that warmer air is normally accumulated in the regions adjacent to that with the low-level anticyclonic circulation, and (2) that this alone results in a tendency to maintain this type of circulation.

2. The Equation for Vertical Velocity by Quasi-geostrophic and Quasi-static Motion

One may write the thermal wind equation,

$$\frac{\delta \mathbf{V}}{\delta z} + \frac{g}{2\Omega_z \theta} \nabla_h \theta \times \mathbf{k} = 0,$$

after taking the vertical component of the curl of each term, essentially as

$$\frac{\delta}{\delta z} \nabla_h \times \mathbf{V} \cdot \mathbf{k} - \frac{g}{2\Omega_z \theta} \nabla_h^2 \theta = 0. \tag{1}$$

(θ = potential temperature)

The change per unit time in the first term due to advection of vorticity and due to friction $\mathbf{F}$ is

$$\frac{\delta}{\delta z} (-\mathbf{V} \cdot \nabla_h \nabla_h \times \mathbf{V} \cdot \mathbf{k} + \nabla_h \times \mathbf{F} \cdot \mathbf{k}),$$

while the corresponding change in the second term due to warming by advection, $-\mathbf{V} \cdot \nabla_h \theta$, and heating Q, is equal to

$$-\frac{g}{2\Omega_z \theta} \nabla_h^2 (-\mathbf{V} \cdot \nabla_h \theta + Q).$$

These two changes generally will not balance each other, which necessarily would lead to a violation of (1) unless compensating changes in the two terms take place as a result of the influences of vertical motion. These are essentially

$$\frac{\delta}{\delta z} \left(2\Omega_z \frac{1}{\rho} \frac{\delta \rho w}{\delta z} \right) \text{ and } -\frac{g}{2\Omega_z \theta} \nabla_h^2 \left(-w \frac{\delta \theta}{\delta z} \right)$$

respectively. To get Eq. (1) satisfied at all times one must accordingly have

$$2\Omega_z \frac{\delta}{\delta z} \left(\frac{1}{\rho} \frac{\delta \rho w}{\delta z} \right) + \frac{g}{2\Omega_z \theta} \nabla_h^2 \left(w \frac{\delta \theta}{\delta z} \right) + h = 0 \tag{2}$$

where

$$h = \frac{\delta}{\delta z} \left(-\mathbf{V} \cdot \nabla_h \nabla_h \times \mathbf{V} \cdot \mathbf{k} + \nabla_h \times \mathbf{F} \cdot \mathbf{k} \right) - \frac{g}{2\Omega_z \theta} \nabla_h^2 \left(-\mathbf{V} \cdot \nabla_h \theta + Q \right). \tag{3}$$

Eq. (2) is the now well-known equation for vertical velocity, w. If h does not have a too irregular distribution with height, it can be shown that the vertical velocity in the mid-troposphere, W, as a first and not altogether bad approximation may be written as:

$$W = K\hat{h}. \tag{4}$$

Here $K > 0$ and a quasi-constant, while $\hat{h}$ is some average value of h in the vertical.

3. The Effect of "Warming" and "Cooling"

Let q denote the warming by advection and radiation,

$$q = -\mathbf{V}\cdot\nabla_h\theta + Q$$

and let W_q denote the corresponding vertical velocity in the mid-troposphere as given by (4), (3),

$$W_q = -\frac{Kg}{2\Omega_z\theta}\nabla_h^2\hat{q}.$$

Since the value of the Laplacian $\nabla_h^2\hat{q}$ at a point, omitting a positive proportionality factor, is obtained roughly as the difference between the mean value of $\hat{q}$ along a small circle centered at the point and the value of $\hat{q}$ itself, it is clear that in regions of maximum and minimum values of $\hat{q}$, $\nabla_h^2\hat{q}$ will have negative and positive signs, respectively. The corresponding sinking motion in, for instance, the region of maximum cooling causes the air to spread at low levels and converge at high levels, producing anticyclonic vorticity at the ground and cyclonic vorticity in upper levels. The low-level winter-time Siberian anticyclone for example is supposed to be created and maintained by such a process in which the cooling is brought about essentially by radiation.

Fig. 2 illustrates in an idealized way how an originally zonal distribution of temperature (I) would change by advection in a simple cellular motion into a distribution (II). Warm air is displaced northwards and sideways, and cold air southwards and sideways. During these changes the maximum advective warming and cooling will be found in the regions marked WW and CC, respectively.

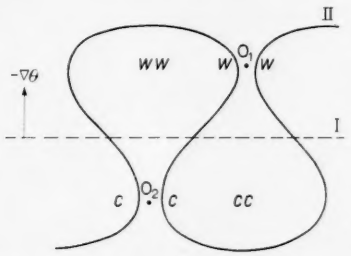


Fig. 2. Illustration of a deformation by advection of an originally zonal distribution of temperature. WW: Region of maximum warming by advection and region of relatively weak and widespread rising motions. CC: Region of maximum cooling by advection and region of relatively weak and widespread sinking motions. O_1: Region of accumulation of warmer air in the close surroundings. Locally strong sinking motion. O_2: Region of accumulation of colder air in the close surroundings. Locally strong rising motion

Besides these regions the ones in the neighbourhood of the points marked O_1 and O_2 are of particular interest. At these points themselves, the advection is small or zero. However, in the first region warmer air has accumulated in the close surroundings of O_1, while the cold air simultaneously has been moved away from the region. Similarly, in the other region colder air is accumulating in the close surroundings of O_2 while the warm air has left the region. During the transition from I to II it is therefore obvious that $\nabla_h^2(-\mathbf{V}\cdot\nabla_h\theta)$ must be positive in O_1 and negative in O_2, corresponding to

sinking and rising motions, respectively. These regions appear as ones where a considerable crowding of the isotherms are taking place. Therefore, in a located region, $\nabla_h^2(-\mathbf{V}\cdot\nabla_h\theta)$ may assume larger values at O_1 than at CC, leading locally to a stronger descending motion than in the more widespread area of sinking motion to the south. Because of the additional warming due to sinking motion in a stable atmosphere this will have as a consequence a tendency to close the isotherms around the cold dome. Similarly, because of the strong ascending motions in O_2 the isotherms will have a tendency to close around the warm dome.

4. The Fields of Advection of Thickness as Obtained from the Normal Flow for November

The full lines in *Fig. 3* show the normal 700–1000 mb thickness pattern for November. The pattern gives the impression of a possible kinematically conditioned squeezing of the isotherms around a point close to the North Pole. The deviation from a purely zonal distribution of the isotherms might of course also be due to a typical distribution of heat and cold sources, and is certainly in part so. To illuminate this question somewhat further the field of advection of thickness as obtained from the normal maps has been computed. The field has been drawn in *Fig. 3* as the stippled lines, the heavy

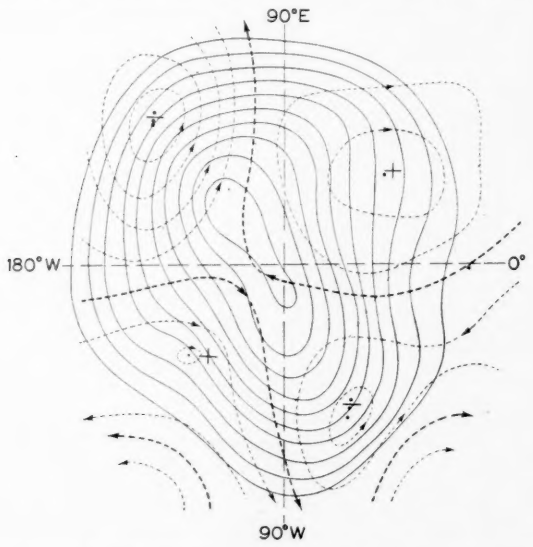


Fig. 3. (a) Full lines give the normal 700–1000 mb thickness lines for November. (b) Stippled lines are isolines of advection. + indicates centers of maximum warming. ÷ indicates centers of maximum cooling. Heavy stippled line is line of zero advection

stippled line being the zero line, and the centers of advective warming and cooling having been marked by + and ÷, respectively. It is seen that if the thickness changed only as a result of this advection, the isotherms would be squeezed still more together, while at the same time the two major cold troughs would be extended farther south. This might be considered as

quite strong evidence for the assumption that the isolines of thickness normally deviate from a zonal distribution because of the kinematic influences from the normal flow pattern which itself deviates in a typical way from a purely zonal flow and which therefore correspondingly would deform by advection any zonal distribution of thickness in some typical way.

5. The Field of Vertical Motion as Obtained from the Normal Flow for November

The distribution of advection as found above for a point close to the North Pole is similar to the one described for point O_1 in *Fig. 2*, and would alone give a tendency for sinking motion in this region. It is an interesting fact that this region coincides quite closely with the central region of the surface anticyclone in the Arctic. However, the Laplacian of advection of temperature is only one of two terms (when friction and heating are neglected) which by Eqs. (1) and (2) determine the regions of ascending and descending motions. Even though the other term is small in the considered region it should be taken account of. On the basis of the normal map for November a vertical average of h

$$-\hat{h} = \frac{\delta}{\delta z}\,\mathbf{V}\cdot\nabla_h\nabla_h\times\mathbf{V}\cdot\mathbf{k} + \frac{g}{2\Omega_z\theta}\,\nabla_h^2(-\mathbf{V}\cdot\nabla_h^2\theta)$$

has been computed whereby the field given in *Fig. 4(a)* was obtained. The positive areas correspond by Eq. (4) to regions with sinking motion and corresponding surface divergence, while the negative areas correspond to regions with ascending motions and corresponding surface convergence. By comparison with the normal surface pressure pattern given in *Fig. 4(b)* it is seen that the computed region of horizontal divergence in the Arctic

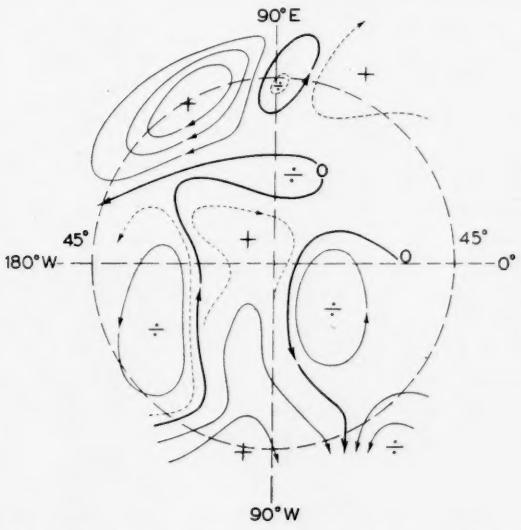


Fig. 4 (a). Field of horizontal divergence at sea level computed from the normal flow of November

Suppl.
Part. I
1956

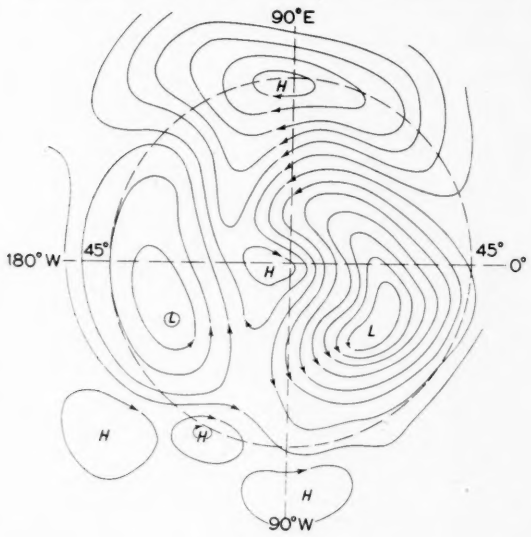


Fig. 4 (b). Normal surface pressure pattern for November

coincides quite well with the region of observed anticyclonic vorticity. It is further seen that there is fair coincidence between the computed areas of surface convergence and the observed regions of cyclonic circulations in the north-eastern Atlantic and Pacific Oceans. However, having neglected such important factors as topography, radiation and friction one would be surprised if regions did not exist where the computed areas of sea-level divergence and convergence disagree with the observed pictures. In this connection it is of interest to point out that a relatively small area of convergence has been computed in the central region of the Siberian anticyclone.

6. Concluding Remarks

In this note the equations of motion have been applied to the normal flow. This can of course not be done in a strict sense without taking into account the Reynold's stress term which appears in the equations when they are smoothed in time. However, the result obtained in this investigation depends entirely upon the fact that the normal patterns of velocity and temperature deviate from a purely zonal distribution, since otherwise the effects investigated would have been identically zero. The only reason for the patterns in question to deviate from purely zonal ones must ultimately be sought in the inhomogeneities of the earth's surface such as the actual distribution of continents and oceans, mountains and so on. The influences from these inhomogeneities are, however, present in each particular situation and might conceivably constitute the most important effects also for more short-range changes, at least on an average. If this is the case, one may be allowed to state that the result obtained would be similar to the average of results btained on the basis of a number of actual situations.

DISCUSSION, 3 JULY 1956

Chairman: F. KENNETH HARE, Canada

KEEGAN: I have a few comments to make. The first one is directed to Mr. Namias concerning increased cyclonic activity in the vicinity of Novaya Zemlya. I have noticed repeatedly that there seems to be a relationship between unusually strong blocking highs in the eastern Atlantic, that is in the vicinity of the British Isles, and cyclogenesis in the Novaya Zemlya region. Your departures from normal showed increased cyclonic intensity in the Novaya Zemlya area, but your positive anomaly occurred farther west in the Atlantic in the vicinity of Greenland. The pressure in the eastern Atlantic agreed with the 1952 normals, so I would think that the 1952 normals were incorrect in the vicinity of Novaya Zemlya and that the last 8 years of data represent a better mean.

NAMIAS: The tele-connection that I described was a clear-cut connection between the mid-tropospheric pressure in the Davis Strait area and that in the vicinity of the Scandinavian peninsula and in Central Europe. When there are positive anomalies, in other words pressure in the Davis Strait is above normal, there is a tendency for negative anomalies in the European area. I believe that this interconnection, when viewed on a hemispheric basis, is an indication that the fluctuation of the past eight years is really a transitory climatic fluctuation, and therefore we should return to a more normal condition as given by the earlier maps.

KEEGAN: My point is that an increased cyclonic intensity in the Novaya Zemlya area would probably indicate stronger blocking action in the vicinity of the British Isles over the past 8 years than over the period of the 40-year normal. If the blocking frequency has been the same during the past 8 years as it was for the years used for the normal maps, then the pressure in the Novaya Zemlya area as represented on the 1952 normal maps is probably incorrect.

HARE: The period 1948–55 is rather longer than the period of ice-islands operation. In this sense, then, that period is not homogeneous as far as the reliability of data is concerned over the innermost pack-ice zone. I have had occasion to compare side by side the analysis over the inner pack, the biggest hole in the observation network, on several different chart series: Deutscher Wetterdienst, the analyses made at Edmonton by the Arctic Forecast Team, and that by the Extended Forecasting Section in Washington; it is quite clear that the measure of disagreement between these maps is a function of whether there are ice-island observations or not in the centre. I suspect that we are in a phase of reaction now against the polar anticyclone, so that people are afraid to draw very high isobars over the inner pack. Quite clearly those islands are of critical importance. I happen to know that Mr. Keegan has made some estimates of exactly how valuable they are in the estimation of pressure in the central zone. I wonder if he would care to make comments on this question.

KEEGAN: In the course of the project I am working on I had to use operational weather maps because the period of study was not covered by the Historical Map Series. To ensure that I could get the best analysis possible, I used two sets of maps: the *Täglicher Wetterbericht* which I assume would be more accurate in the Eastern Hemisphere, and the *Daily Northern Hemisphere Charts* prepared by the Extended Forecasting Section of the U.S. Weather Bureau, which should have more data

Suppl.
Part.I
1956

from the Western Hemisphere. Essentially there was one difference between the two sets of maps. The German weather maps did not have, or did not use, the data from T-3. I decided to compute the difference in the actual pressure at T-3 and the pressure as analysed without the benefit of this weather report. For the month of December 1953 the *Täglicher Wetterbericht* reported an average pressure 7 or 8 mb higher than the pressure actually observed at T-3. This error ranged from +17 to −2 mb. There were only two cases in which the German maps reported an analysed pressure lower than the one reported at T-3.

NAMIAS: I think that there is no question that researchers dealing with day-to-day analysis in the Arctic are bound to have difficulties and problems for many years to come, unless the network is improved considerably. There are break-away systems, of which I spoke this morning, that are sometimes of rather small scale, difficult to catch. There are also break-away anticyclones from the Siberian high, which come across the Polar Basin and which also are sometimes difficult to detect, leading to errors in day-to-day maps. But when one computes means from twice daily maps for a long period of time, the probability of large errors is diminished considerably. The point I wish to make is that I think we have, for the lower troposphere, now obtained a reasonably reliable normal pattern upon which we may base estimates. It appears certainly good enough to evaluate anomalies which arise in connection with predictions for 5, 10, 15 or 30 days. That is, anomalies for these periods are large enough to be easily detected and detailed, using as a background the present normals. It is even difficult to assess the differences between the two sets of normals I have presented in terms of transitory climatic fluctuations or changes in the normal estimates. I do not think the problems involving mean fluctuations are anywhere near as serious as those of resolving the differences between analyses of day-to-day charts.

I should also like to make a point in connection with judging the probable correctness of above or below normal pressures in the Arctic. If maps for the summer of 1955 are constructed for 5-day periods or even monthly periods, we find that practically every period was below normal over the Polar Basin. One might first come to the conclusion that the normals were wrong by some amount. But this conclusion is erroneous, for this persistence is quite characteristic of mean maps. For example, during the summer 1955 the circumpolar vortex was unusually contracted. We cannot say that just because a period is characterized by an anomaly of consistent sign the normal is wrong by a certain amount. I doubt very much that subsequent years of data are going to change the Arctic sea-level picture of the normal radically, certainly not to the extent of 5 or 10 mb.

SCHERHAG: I believe it will be very difficult to solve such problems as are raised now. First I will agree with what Dr. Hare suggested, that some of these central values of the highs in the polar regions are estimated too low.

With respect to the different operational analyses, some of the big discrepancies in the Arctic may be due to different sources of errors in the transmission of the Russian data.

As to the circulation in the Arctic in the year 1953, which has been mentioned frequently here today, we have investigated this circulation in Berlin, and we found that during this whole year the pressure values were very low in the entire Arctic region. In every month we had pressure deficits in the Arctic region of about 5 to 10 mb. And even in the yearly average for 1953 there was a deficit of about, I believe, 8 mb. This was a year which showed some special form of the Arctic circulation. It was a year which was especially warm in the stratosphere, even in February.

Then I should like to make some remarks with respect to the changing of the whole circulation between February and March. I would believe, in order to solve

this problem, we must at first look at the maps showing the pressure variations from month to month. In Berlin we are now constructing new normal maps and also the changes from month to month. On these maps we see first that between February and March the Icelandic and Aleutian lows fill. And the question is, why does this low fill between February and March? But we not only have pressure rise in the Icelandic and Aleutian lows. We have a big pressure rise over the entire polar region, so that we have a pressure rise in the entire area north of 55 or 60 degrees latitude. And we have a pressure fall in the middle latitudes. In my opinion this must be associated with the fact that the temperature differences are no longer so strong between the cold mass over North America and the Atlantic Ocean, so that the corresponding frontal zone becomes weaker from February to March. The filling up of the Icelandic low begins even in February. At the same time the polar region becomes the coldest region of the Northern Hemisphere, and it has the highest pressure in March and April. And it might be that this causes the breakdown of the whole general circulation. But I believe we cannot solve these problems now.

SCHMIDT: I should like to ask Dr. Flohn a question. He said, as I understood, that there is a difference in the diurnal variation of temperature between high pressures over Europe and over Siberia. And I think he connected this difference in the diurnal variation of temperature with the rate of subsidence in both anticyclones. I should like to comment that it is perhaps more probable that the greater diurnal variation in temperature occurring with anticyclones over Europe is due to the fact that slight variations in wind direction immediately give rise to other temperatures because of the seas surrounding Europe, an effect that is not present with Siberian anticyclones.

Suppl.
Part.I
1956

FLOHN: I agree with this remark. To exclude misunderstandings I should mention that the standard deviation of temperature was derived for all weather situations. If the standard deviation of temperature is increasing with height, which is normally the case in Europe and apparently in large parts of middle latitudes, this is in most cases due to advection. But if the standard deviation of temperature is decreasing with height, as near the Siberian and Canadian cold pole during winter, advection of tropospheric air-masses cannot explain the temperature variation from one day to another. It can explain only the temperature variations in the lowest layers, below the inversion. There must be a dynamical effect, or may be a combination of thermal and dynamical effects, that produces such a remarkable constancy of upper-air temperatures in the Siberian and in the Canadian cold pole during winter.

KEEGAN: And then my comments to the paper presented by Mr. Moreland. I have been working with winds from 1949 to about 1955 over the polar regions. The year 1953 was the only one in which the winds at 50 and 100 mb were either weak westerly or in some cases easterly in the mean for the month of January. I think it was unfortunate that the year you had to analyse disagreed with Kochanski's and Wasko's maps. In the summer I agree with you that westerly winds exist up to 20 km, but they are weakening at a fairly rapid rate. At 30 km there are, in general, easterly winds. I think Kochanski's work is qualitatively correct, but I disagree quantitatively with the strength of both the winter jet stream at the 50-mb level and the summer easterlies.

MORELAND: I agree with Mr. Keegan that 1953 was rather an anomalous year at these high levels, especially over North America. However, over the European area we found a persistent intense cold low, which also, no doubt, was an anomaly for this area. Undoubtedly, there was quite a strong shift in the stratosphere from the normal patterns. I would agree also that during the summer the wind flow becomes easterly at 30 km in the vicinity of the Pole. Our summer synoptic 50-mb charts indicate that periods of easterly winds may occur in this region even as low as the 20-km level. These are caused by the stagnation of a high-pressure area in the

vicinity of the Pole. We find that at 50 mb quite wide variations may occur from year to year.

SUTCLIFFE: I would like to ask Dr. Godson what is the degree of coupling between the oscillations in his high-level jet and the oscillations in the lower troposphere. We know that over a large part of the world the general run of the contours of 100 mb follows fairly well the contours in the lower levels. At 300 mb they tend to be roughly parallel, they do not give the impression of independent instability waves.

So I would like to know whether he believes that the oscillations in the stratospheric jet are induced spontaneously within that high baroclinic zone, which is unstable dynamically in itself, or whether they are merely forced oscillations induced by the baroclinic oscillations which we know exist in the lower atmosphere. I am rather interested in this because in the meteorological literature, although I have read quite a lot about instability in baroclinic zones, I have never seen a study of the baroclinic instabilities which could occur in the upper atmosphere. This is important, not only in connection with the stratospheric jet, but in connection with the subtropical jet. The subtropical jet we now know has practically no baroclinity up to 500 or even 300 mb, and then a very large baroclinity up above. And I think it is now generally agreed that the subtropical jet is basically distinct from the polar-front baroclinic zone and the strongly oscillating middle latitude jet. These are two physically distinct phenomena, although unfortunately in North America at 80°W the frontal jet in the winter is so far south that it gets mixed up with the subtropical jet. I wonder if Dr. Godson can tell me whether he has shown synoptically that the oscillations are independent and can be quite out of phase with the surface oscillations and whether he has obtained a dynamical criterion for the instability of a rather shallow baroclinic zone embedded in a rather deep layer of limited baroclinity.

GODSON: In the first place, we have not yet made any studies with these data relating the events in the midstratosphere to specific phenomena in the troposphere. However, studies of mean temperatures throughout the winter showed that the north-south temperature gradient increased with height in the stratosphere. Examination of the cross-sections constructed showed no evident coupling of the stratospheric jet and baroclinic zone with tropospheric conditions. Finally, the reference to the baroclinic instability was intended primarily as an apt qualitative description of the phenomenon. We have not yet treated the problem quantitatively on this basis, although dynamic instability criteria derived from the parcel approach are, in theory at least, applicable to isolated layers of intense baroclinity.

NAMIAS: I have a few remarks dealing with the same general problem of a coupling mechanism; I would like to ask if there is someone in the audience who might have related ideas. In three papers delivered today associated phenomena were discussed.

The first was that by Wexler and Moreland who spoke about some of the factors possibly leading to a strong increase in ozone, at a certain time of the year. The second was the paper presented by Dr. Godson who told us about the high-level jet, and the large variability associated with the migration northward of this jet. The third was my own paper in which I spoke briefly about the index cycle which, though observed primarily in the lower troposphere, there is every reason for suspecting is a very deep phenomenon of the atmosphere. Now studies of the index cycle of the lower troposphere suggest that there is a preferential time of the year for this cycle to occur. Even the climatological means of some 9–10 years of data show an index cycle with a minimum of the westerlies occurring around the end of February or first part of March, and then increasing. This appears to be the "primary" index cycle of the year. There is a secondary period around November during which such cycles are frequent. One has a feeling that these items, the ozone, the polar-night jet, and the index cycle are all associated phenomena. If they are, then the problems of interdependence become very vital from the standpoint of long-range weather

prediction. In other words, delays in the index cycle, its character, duration and intensity, and variations in ozone, in the migration of the polar-night jet may be reflections of the same general phenomenon. I raise this possibility to suggest research along these lines and also to ask if there is someone in the audience who has more definite ideas about the possible coupling than I.

SUTCLIFFE: I think it was mentioned by Mr. Moreland that the rather rapid increase of ozone was probably due to the breakdown of the jet and the turbulent transfer of the ozone. I am wondering whether the meridional overturning has been computed. During the phase of creating a new solenoidal distribution there must be overturning. That might be quite a large term in the sudden increase of ozone, so that you could get quite a considerable transfer of ozone northwards, not by turbulent mechanism, but by meridional overturning around the increasing jet.

I think it would be unfortunate to close the meeting without comment on what I think is quite an important paper presented by Dr. Fjørtoft. Our chairman has remarked that the polar anticyclone is getting a bit unfashionable. I wonder if that is because he is shaken by the lack of theories. But now we have one from Dr. Fjørtoft. It seems then that at least one term, one important term, which will account for the stability of the extremely asymmetrical circumpolar circulation, with its two cyclones and two anticyclones, and the anticyclonic ridge over the Pole, may be provided not by Reynold stresses, not by turbulent diffusion, but by a steady state solution. If there is a steady state solution to get the transfer of vorticity or momentum or whatever parameter you like, then I think that it is really rather important, and I do not think it is at all familiar in the meteorological literature. At the present time every one is putting a great deal of weight on the Reynold stresses. And I would ask Dr. Fjørtoft how far he has carried out this quantitative work, whether he has shown that these, what are effectively steady state terms, are adequate quantitatively to provide for the loss of vorticity by friction at the surface.

FJØRTOFT: I must admit that I have thought of just what Dr. Sutcliffe mentioned. It might be possible to find, taking the topographical influence and so on into account, steady state solutions which are essentially of the character Dr. Sutcliffe mentioned. But I think it will be impossible then to isolate the problem to the polar region. For instance, I think I showed today that the sinking motion in polar regions resulted from the accumulations of warmer air in the surroundings which in turn were due to the deviations in the normal pattern from a pure zonal flow farther to the south. So you have to consider a large part of the Northern Hemisphere. Such steady state solutions in a very simple way have already been found in cases where it is possible to linearize the equations, as, for instance, shown in Charney and Eliassen's work and in Smagorinsky's work. The effect I have introduced is, however, a typical nonlinear one, and it would indeed be a difficult task to extend the steady type solutions to the nonlinear case.

Supp
Part
1956

SECTION III

uppl.
art.I
956

Sup
Par
195

THE ROLE OF TROPOSPHERIC COLD-AIR POLES AND OF STRATOSPHERIC HIGH-PRESSURE CENTERS IN THE ARCTIC WEATHER

R. SCHERHAG
Berlin

Suppl.
Part. I
1956

THE Meteorological Institute of the Free University of Berlin has devoted in the last 3 years some work on the conditions of the Arctic atmosphere, especially in the regions of cold-air poles and stratospheric high-pressure centers. A few dissertations dealing with this subject will be published soon, and I shall refer to those problems only which are of more general interest.

The main research project sponsored by American funds deals with the general circulation and there have been constructed new normal surface and upper-air charts for the Northern Hemisphere, covering the period 1900–39. These charts refer to the 1000, 850, 700, 500, 300, 200, 100 and 50-mb surface, respectively. They are constructed by reducing the means of the period 1949–56 to the 40-year means, and the material available was completed by calculating the monthly averages for about 100 Russian and Siberian radiosonde stations whose results were received by radio at the German Weather Center at Frankfurt in the years 1949–53. Great care was used to eliminate the differences between the individual radiosonde types and the differences resulting from radiation errors.

The map showing the yearly averages of the height of the 1000-mb contours (*Fig. 1*)[1] may give an idea of the features dominating the circulation of the Northern Hemisphere. The general circulation is governed by three anticyclones, which are situated, in the yearly average, over the Azores region, the eastern Pacific and central Siberia at sea level and whose axes are tilted to the south-west. At the 500-mb surface (*Fig. 2*) the high-pressure centers coincide closely with the warmest regions of the world: Central America and the Caribbean Sea, the western Pacific east of the Marshal Islands and the region between India and Central Africa.

There are only two trough-lines situated along the eastern territories of North America and Asia. The only wedge fairly discernible is that extending near the western coast of Europe. The other wedge, supposed to be situated near the Ural Mountains, is not confirmed, and of the third wedge supposed to be situated about the Rockies is only left a weak remnant over Alaska.

Both the trough-lines and the wedges remain nearly stationary throughout the year. In winter, however (*Fig. 3*), there are two pronounced cold lows situated over eastern Siberia and the Canadian Archipelago. The centers of the warm anticyclones hold the same position as in the yearly mean in all heights up to the 50-mb surface (*Fig. 4*),[8] but in this height there seems to be left only one pronounced cold low situated near the North Pole.

Special attention may still be paid to some interesting features of the mean monthly surface pressure changes over the polar region. Thus the pressure rise occurring over the whole polar region from January to February (*Fig. 5*) is worth mentioning and is associated with a widespread pressure fall in temperate latitudes. This accumulation of air in the Arctic region is originated by the warming already beginning in temperate latitudes while temperature in the interior of the Arctic just reaches its lowest value.

This pressure rise in the Arctic region continues until March. In April the pressure rise begins to move to northern Europe, where it will lead to that fine weather generally observed in May (*Fig. 6*). In June (*Fig. 7*), however, a widespread pressure fall begins over the whole Arctic region and continues until July (*Fig. 8*). This pressure fall, according to the most recent observations performed in the interior of the Arctic in the years 1952–56, is revealed to be much stronger than the pressure changes shown in the maps constructed in Berlin in 1953[2] and in the Normal Weather Charts of the U.S. Weather Bureau.[3] This pressure fall, I would suppose, will be originated by radiational heating of the upper atmosphere which in this season is strongest over the polar region. The pressure rise at the same time is moved away to the Aleutian and the Azores anticyclone, respectively. Thus the westerly wind component over the Northern Atlantic and central Europe is strengthened considerably, leading to the so-called European Monsoon.

From July to August the reverse pressure changes begin. The pressure fall in the regions of the Islandic and the Alëutian low in September does not reach the North Pole, and the pressure rise increasing there in November and continuing in December is only partly interrupted in January.

The individual cold poles defined as the main cold centers of the Northern Hemisphere showing 500-mb temperatures below −30°C in the summer season and below −40°C in the other months, are originated by lifting processes in the interior of the Arctic cyclones. This is clearly exhibited in the later part of summer 1953, when temperatures as low as −40°C were already reached in the midst of August instead of at the end of September as in 1954 and 1955. These low temperatures were associated with an unusual cyclone activity.

According to the synoptic weather map of 4 August 1953 (*Fig. 9*)[9] there was situated an unusually deep cyclone with pressure values below 985 mb near the North Pole. Near the center of this cyclone (*Fig. 10*) the 500-mb temperatures are as low as −35°. Ten days later, on 15 (*Fig. 11*) and 16 August (*Fig. 12*), the 500-mb temperatures have decreased even below −40°C, and until the end of September of this year temperatures remained far below the values observed in 1954 and 1955, for instance.

One of the main problems related to the behavior of the cold lows lies in the forecasting possibilities of their path. For this purpose W. Klug has performed, in his dissertation,[4] a special investigation of a series of 32 cold lows, characterized by definite centers at the surface extending nearly perpendicularly through the troposphere.

In *Fig. 13* there are shown the mean tangential wind-speed components around all the 32 cold cyclones travelling in the direction of the arrow plotted on the map. The strongest wind speed, in an eastward running cyclone, for instance, is measured in the south-western quadrant. The wind

Suppl.
Part.I
1956

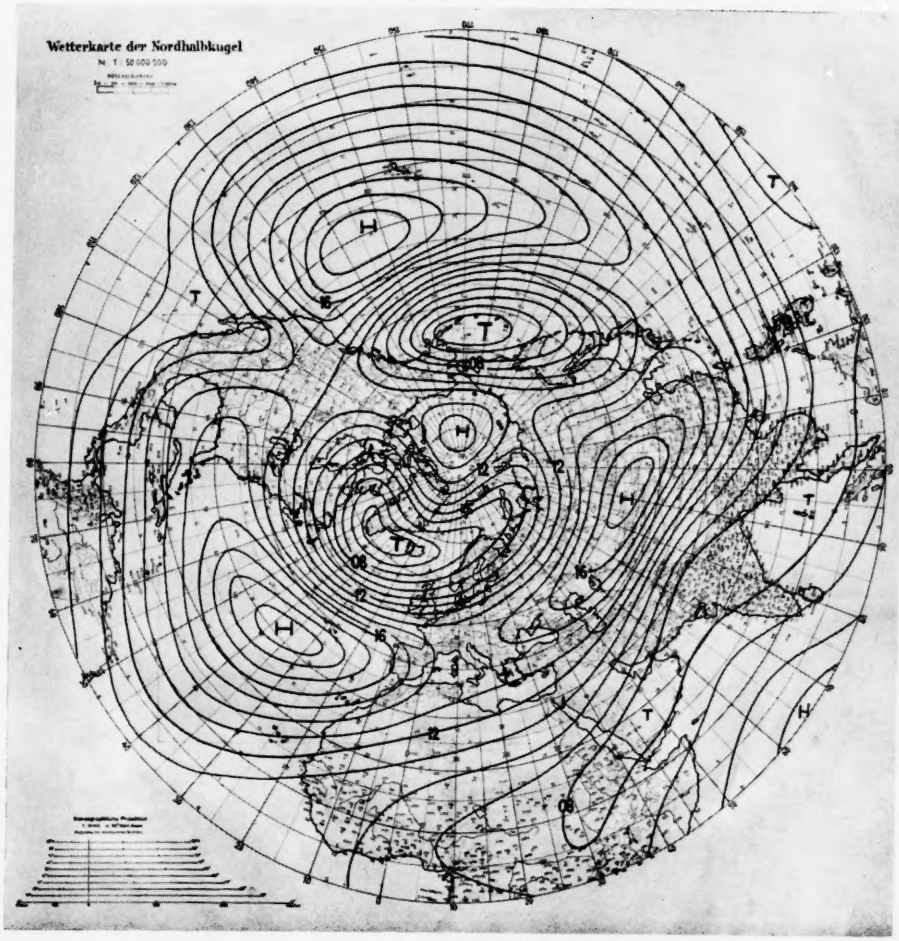


Fig. 1. Yearly averages of the 1000-mb contours

(*Facing page* 102)

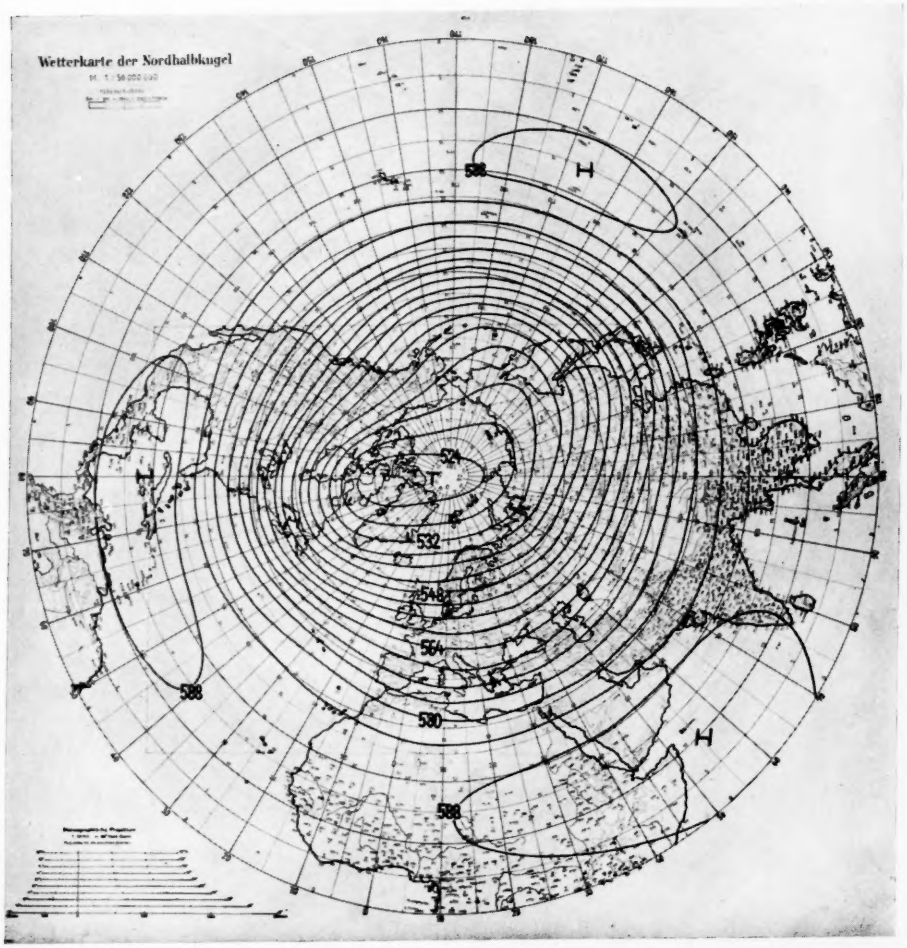


Fig. 2. Yearly averages of the 500-mb contours

Suppl.
Part. I
1956

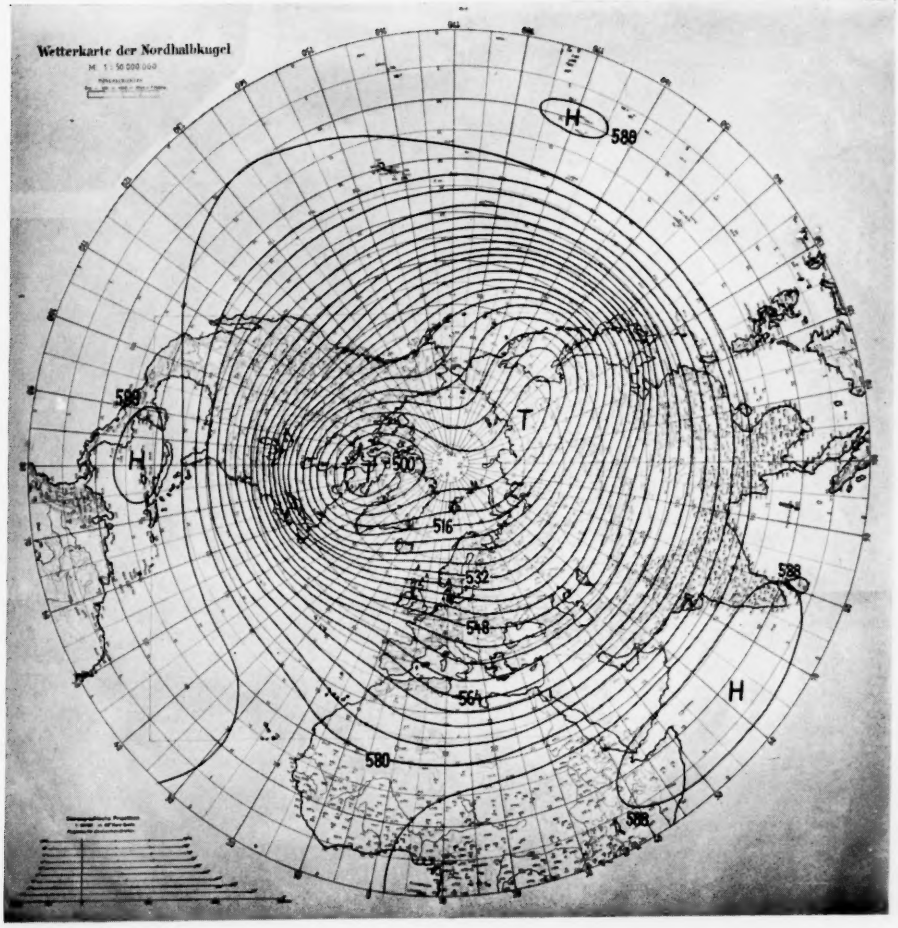


Fig. 3. Mean January 500-mb contours

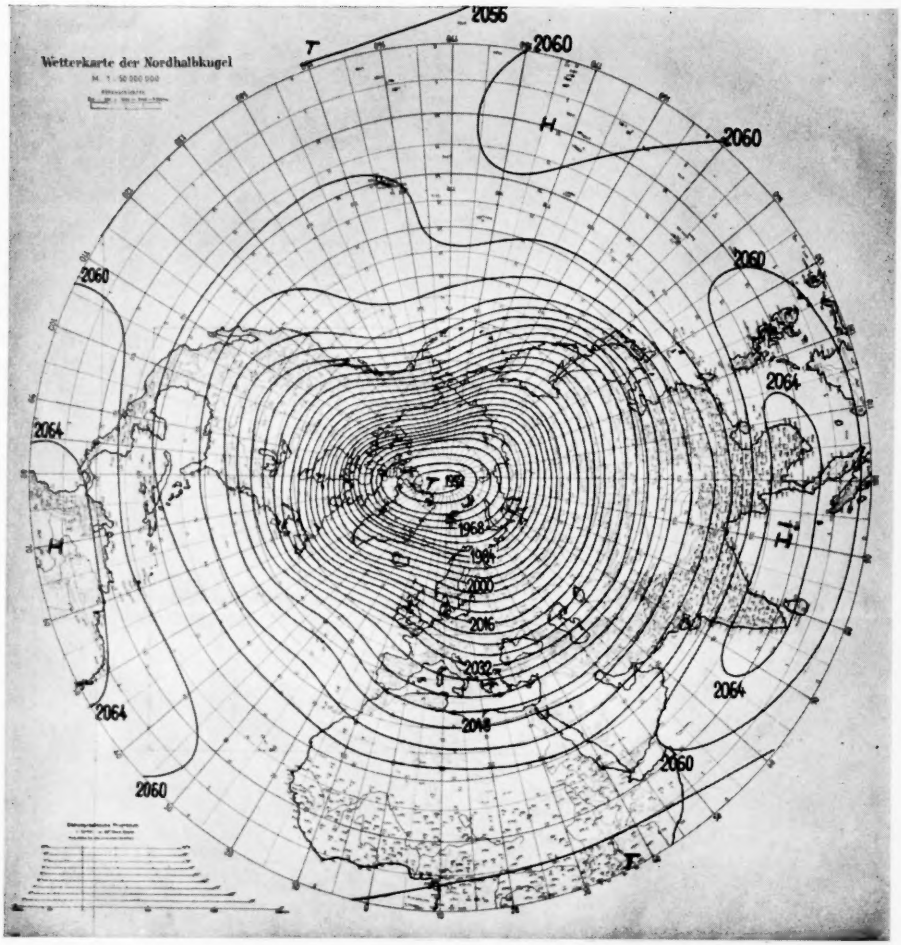


Fig. 4. Mean January 50-mb contours

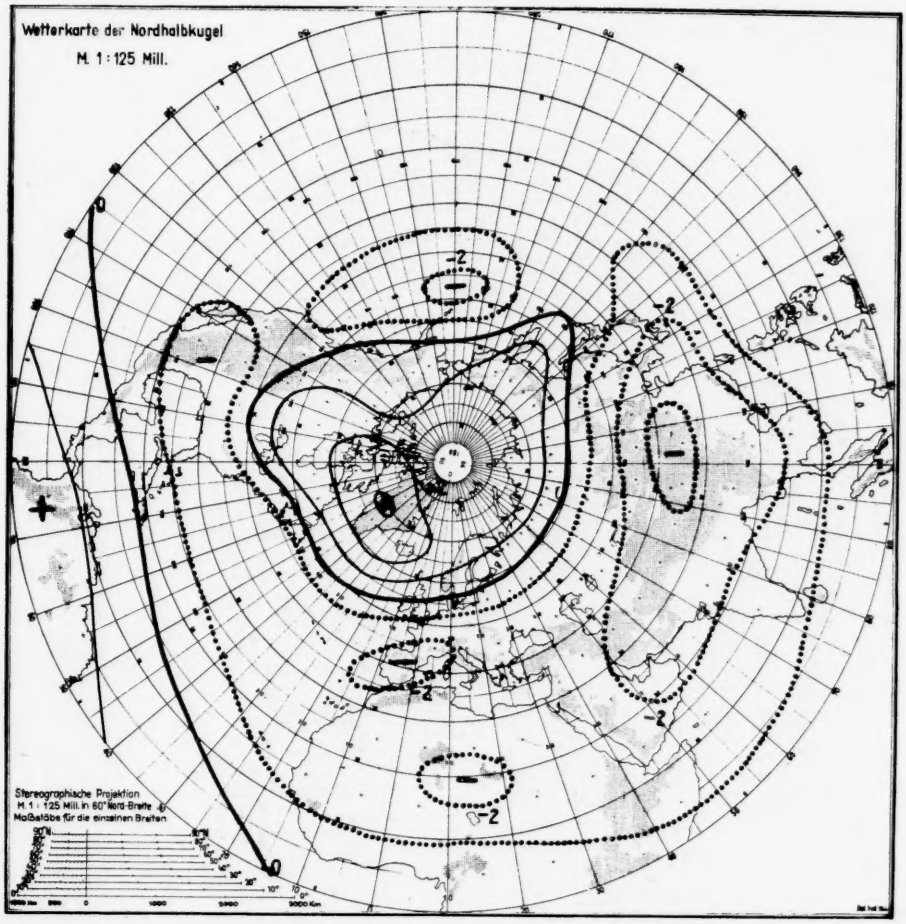

Wetterkarte der Nordhalbkugel
M. 1:125 Mill.
Stereographische Projektion
M 1 : 125 Mill. in 60° Nord-Breite
Maßstäbe für die einzelnen Breiten

Fig. 5. Mean monthly changes of 1000-mb contours from January to February (gpdm)

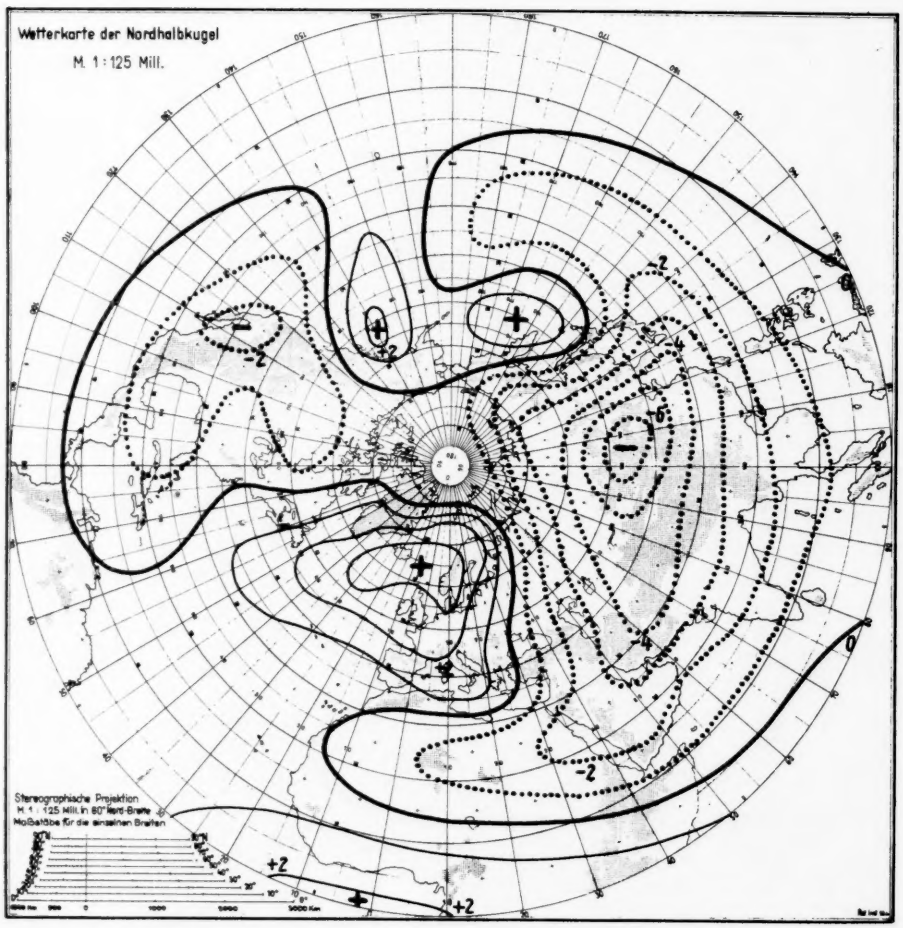


Fig. 6. Mean monthly changes of 1000-mb contours from April to May (gpdm)

Suppl.
Part. I
1956

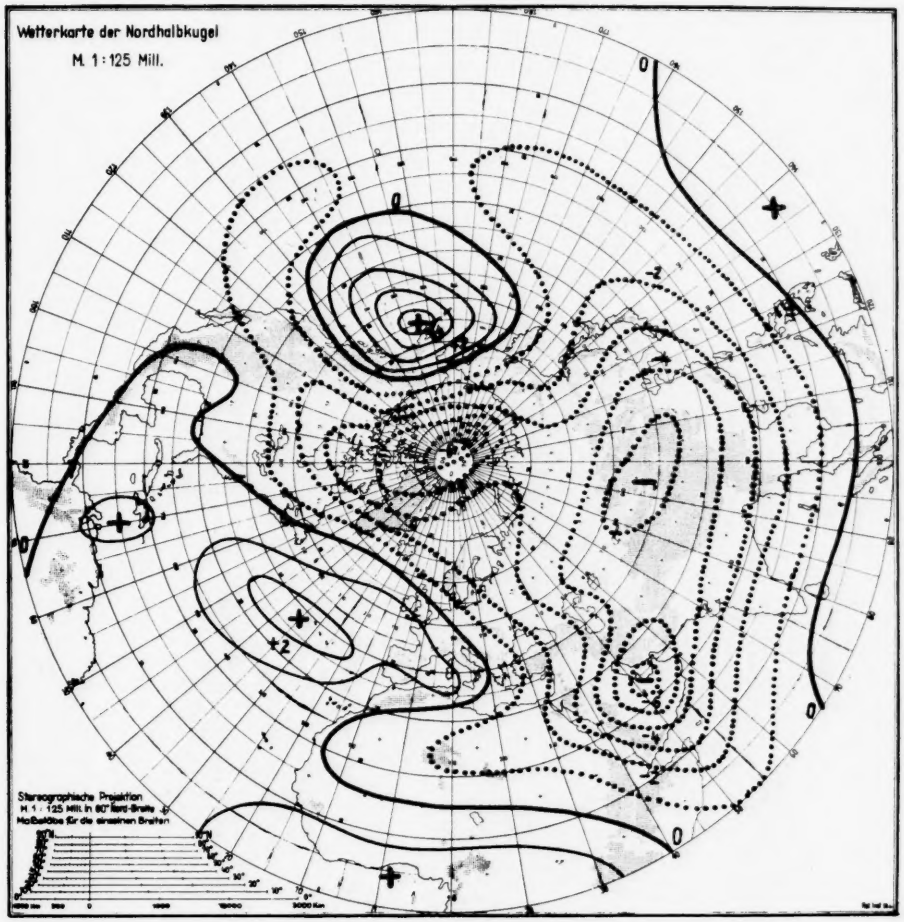


Fig. 7. Mean monthly changes of 1000-mb contours from May to June (gpdm)

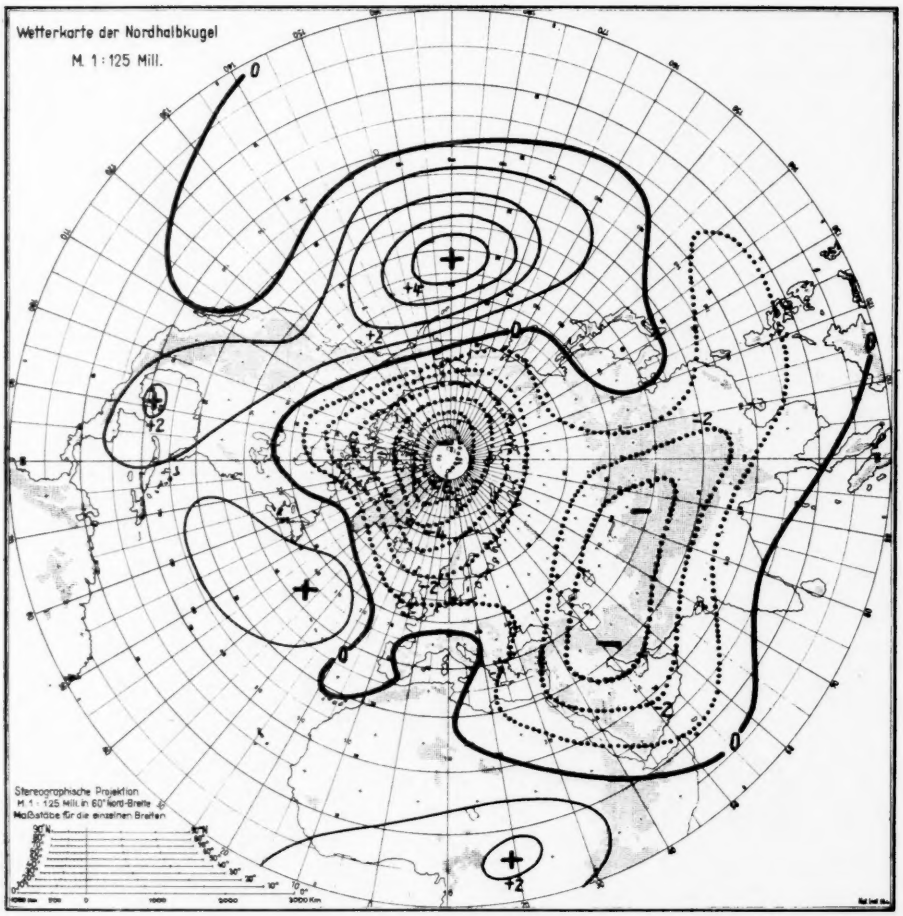


Fig. 8. Mean monthly changes of 1000-mb contours from June to July (gpdm)

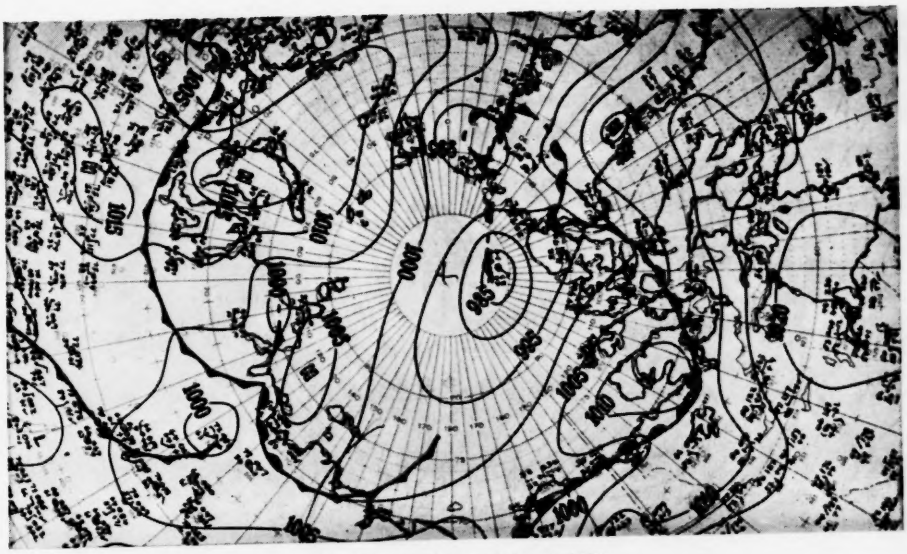


Fig. 9. Sea-level map, 4 August 1953

Suppl.
Part.I
1956

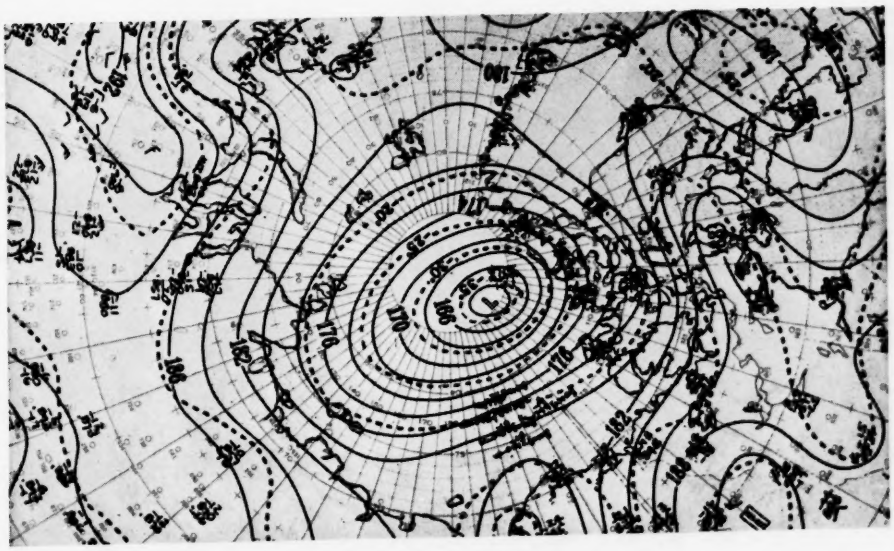


Fig. 10. 500-mb map, 4 August 1953

(*Facing page* 106)

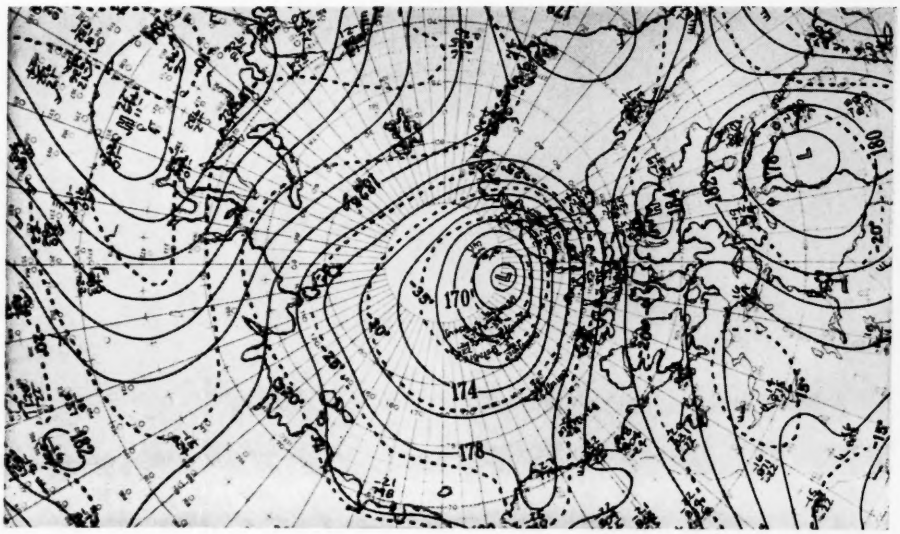


Fig. 11. 500-mb map, 15 August 1953

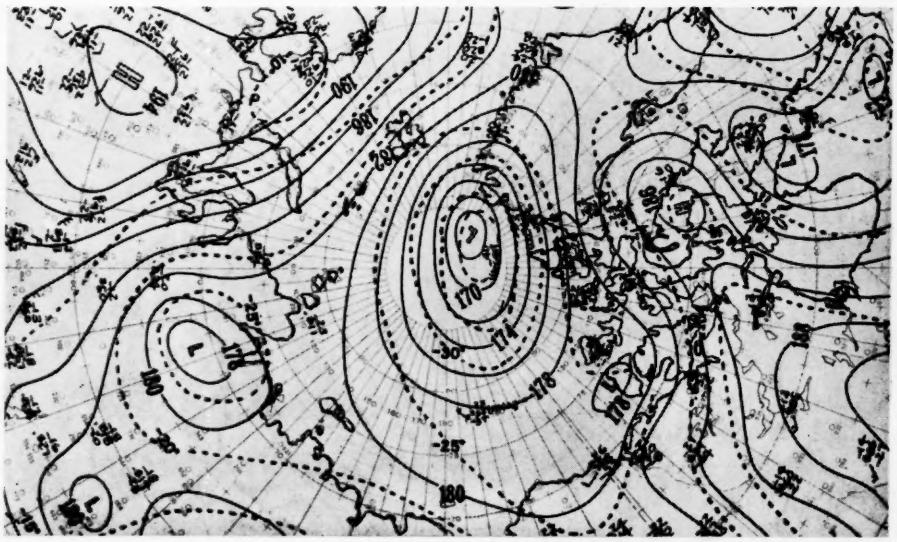


Fig. 12. 500-mb map, 16 August 1953

(*Facing page* 107)

speed decreases rapidly on the foreside of the low. In this region there is observed, according to *Fig. 14* showing the mean 500-mb radial wind-speed components of the 32 cases, a pronounced outflow, and on the rearside a compensating inflow.

The future path of a cold low is given, in the same way as proved by Mrs. E. JORDAN[5] with respect to tropical storms, by the mean resultant wind vector in the environment of the cyclone averaged through all layers. For practical purposes Klug has found it to be sufficient to take the mean vector of all wind vectors measured about 300–500 km around the low center

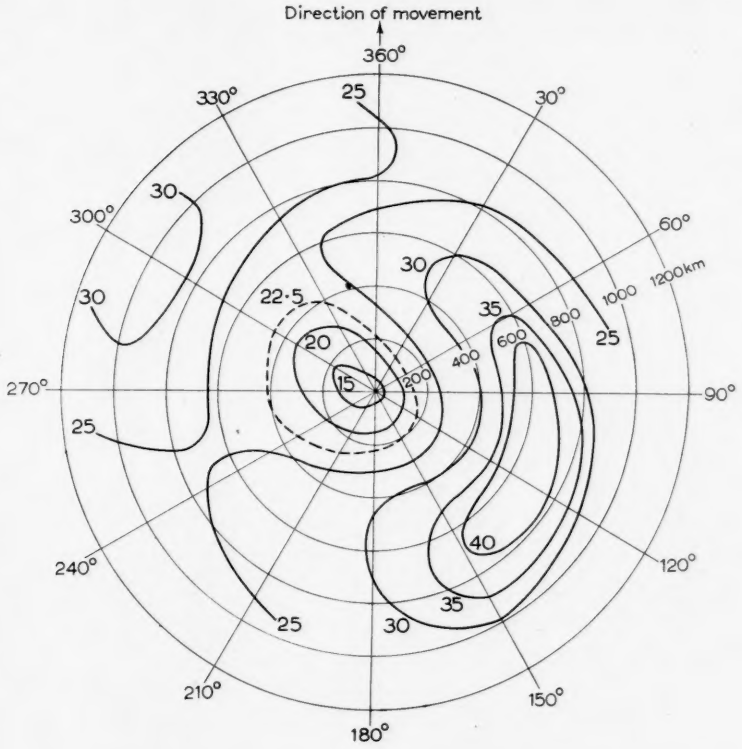


Fig. 13. Mean tangential wind-speed components at the 500-mb level (kn) around a cold low

at the 500-mb surface divided through the number of the vectors used. As an example, according to the *Täglicher Wetterbericht des Deutschen Wetterdienstes*, there may be discussed the weather situation on 21 May 1954, when such a cold low was situated over north-western Germany and it was difficult to decide in which direction it would move. According to the method just mentioned, Klug calculated (*Table 1*), with respect to this cold low, a direction of 304° and a path of 210 km in 12 hours, and really there was observed 290° and 230 km. In the afternoon strongest wind speeds shifted to the north-westerly quadrant of this cold low, and there was calculated a

Suppl.
Part.I
1956

Table 1

Date	*Time*	*Prognosticated path direction and speed*	*Observed path direction and speed*
21 May 1954	0300 z	304° 210 km/12 h	290° 230 km/12 h
21 May 1954	1500 z	259° 105 km/12 h	270° 130 km/12 h
22 May 1954	0300 z	247° 120 km/12 h	240° 140 km/12 h

movement direction of 259° and 247°, respectively, until 22 May. The real path is in close accordance with the calculated values.

In the case of a so-called "cold air drop" defined as a "cold pool" or a "cut off" not extending to the ground, the nearly straight surface isobars represent the prevailing vector in the environment of the system and thus can be taken for prognosticating the future path. An interesting case of this type occurred in the polar region in March and April 1951. Between February and March of this year there occurred a pressure rise over the polar region (*Fig. 15*) of about 20 mb leading to a nearly stationary high-pressure system over the Arctic region (*Fig. 16*).

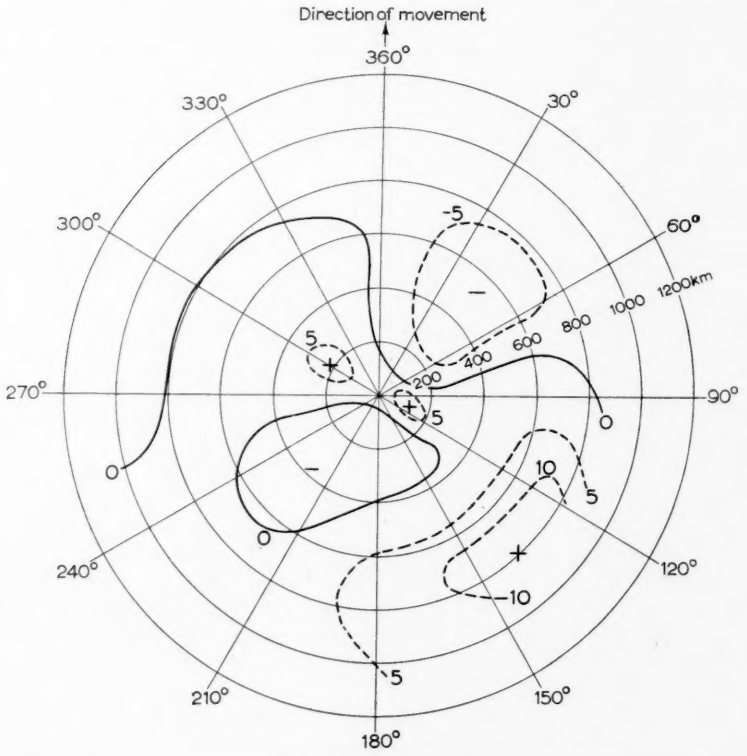


Fig. 14. Mean radial wind-speed components at the 500-mb level (kn)

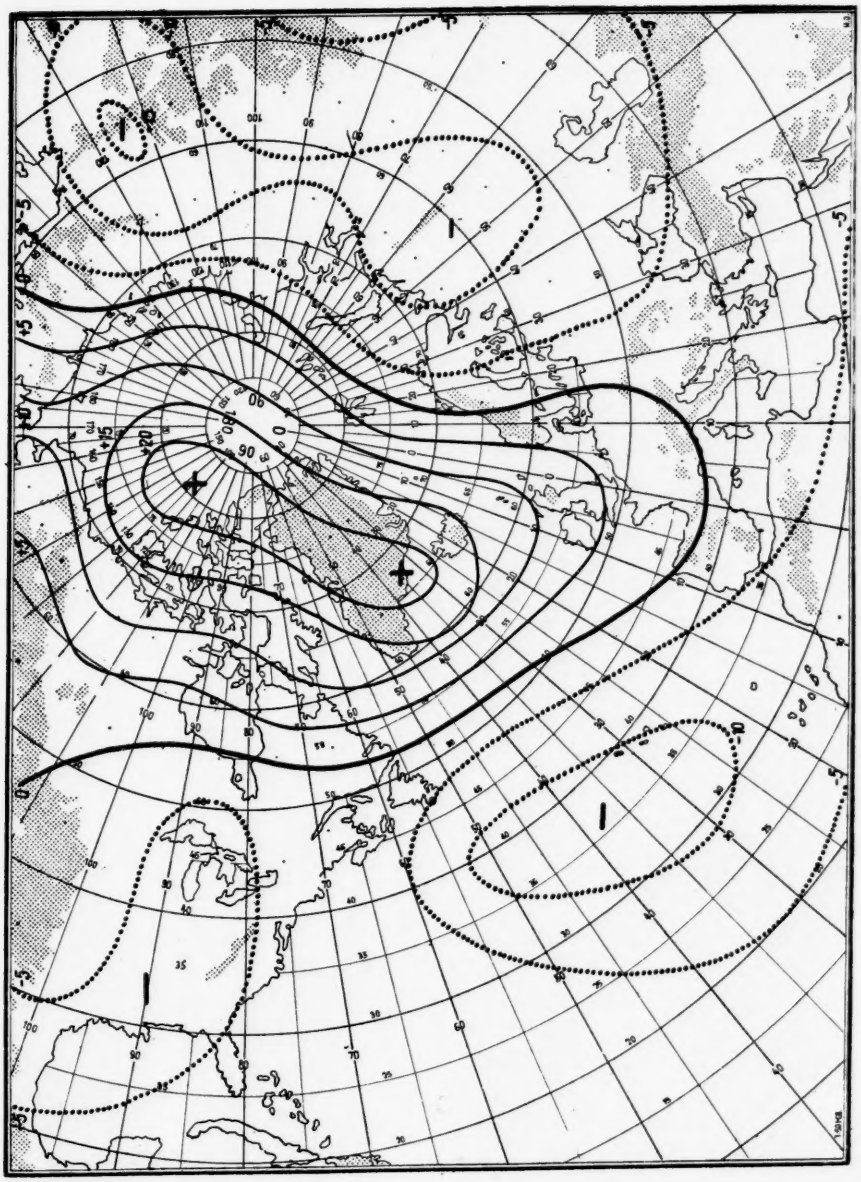


Fig. 15. Pressure changes from February to March 1951 (mb)

uppl.
art.I
956

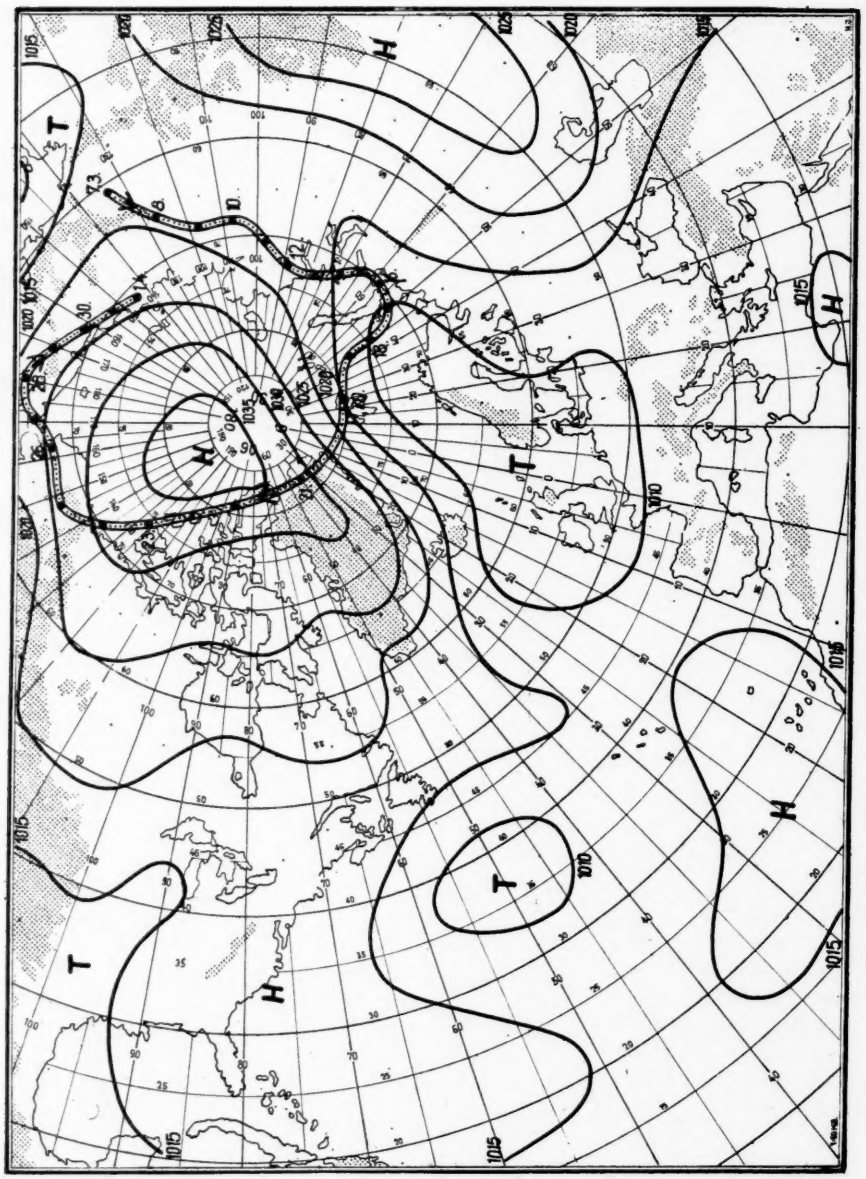


Fig. 16. Mean surface-pressure distribution in March 1951 and simultaneous track of a cold pole

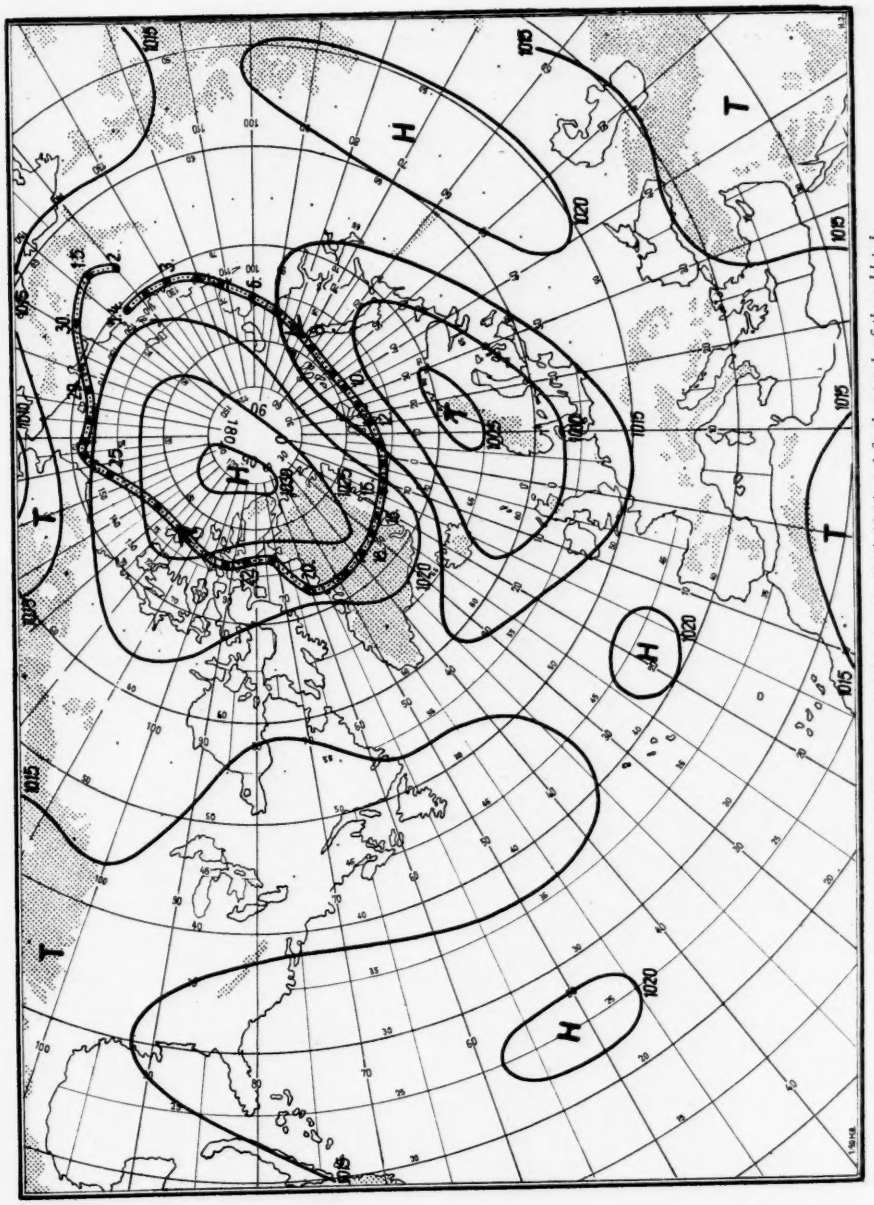


Fig. 17. Mean surface-pressure distribution in April 1951 and further track of the cold pole

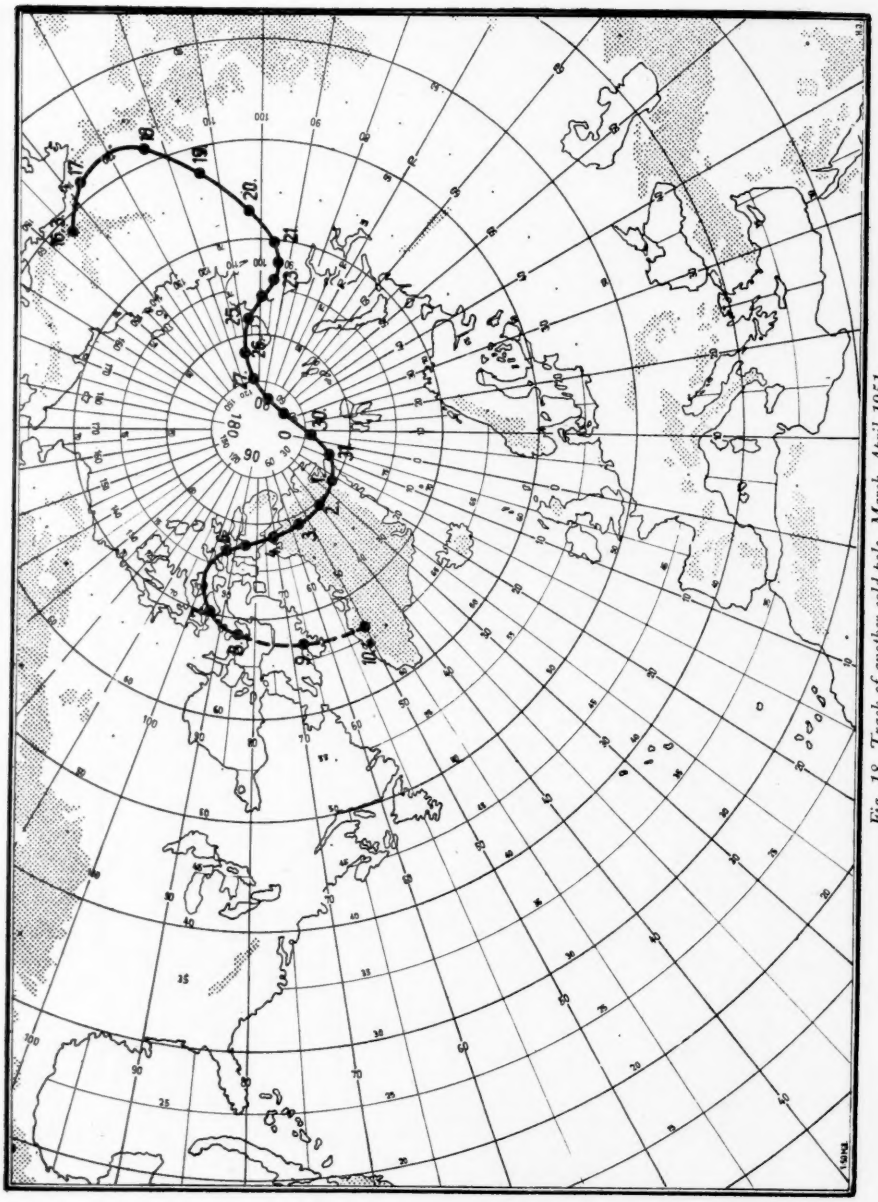


Fig. 18. Track of another cold pole, March–April 1951

A cold-air drop situated on 7 March over eastern Siberia was caught by this anticyclone and was steered, according to the heavy interrupted line in *Fig. 16*, around it in a whole circle until it arrived on 1 April nearly at the same position as three weeks ago. During the next four weeks pressure distribution equally remained nearly stationary, and this cold-air drop continued its path, circling the North Pole once more until it arrived at the same position over north-eastern Siberia again at the beginning of May (*Fig. 17*). Another cold air drop, situated over eastern Siberia on 16 March, took the same way, but left the polar anticyclone over the Canadian Archipelago about the 7 April (*Fig. 18*).

The variation of the height of the 500-mb surface over the whole polar region was governed by these migrating cold-air drops through March, April and May 1951. This is shown by the curves in the right half of *Fig. 19*. The first curve refers to the 500-mb surface values at Jakutsk in the interior of Siberia. The small arrows show the passing of the main tropospheric low which there arrived three times. The crosses refer to the other cold-air drop just referred to. At Cape Tscheljuskin at the northernmost edge of Asia the main cold center passed a few days later, and the curve of Franz-Joseph-Land shows the same singularities. Even the surface-temperature curve of Hopen Island exhibits clearly the passage of this cold polar mass, and it was even recorded by temperature observations at the station "Eismitte" operating in the midst of the Greenlandic Icecape. Over Thule the same phenomena were observed, and in the curve representing the 500-mb variations at Point Barrow in northern Alaska and at the station Chetyre in the eastern Siberian Sea the passage of the cold pole is even best pronounced.

Suppl.
Part.I
1956

The curves on the left side, representing the pressure or temperature variations averaged for a period of 30 days, exhibit clearly the overwhelming influence of this Arctic cold pole on the Arctic weather in this period. The heavy line combining the lowest values of the individual curves, shows the gradual delay in the arrival of the lowest upper-air pressures or surface temperatures, respectively, and thus proves the steady migration of the main cold pole around the Arctic region.

It is not only the Arctic troposphere which offers many problems related with closed centers of low or high pressure, respectively, but also the Arctic stratosphere shows some interesting peculiarities. Thus the dissertation prepared by G. WARNECKE[6] deals with sudden stratospheric pressure and temperature rises and the associated establishment of stratospheric high-pressure centers.

In *Fig. 20* there are represented the temperature curves 500 mb through 50 mb over Alert between November 1951 and 31 March 1952. Whereas the surface temperature rise occurring in the last decade of January 1952 did not extend far into the stratosphere, the unusually strong stratospheric temperature rise of February, on the other hand, was scarcely pronounced in the lower troposphere.

Such a phenomenon of stratospheric heating as observed in winter 1951–52 did not happen in the following year, but was observed again, though not so well pronounced, in March 1954. This is shown by *Fig. 21* containing the 100-mb temperatures observed over Alert in those three winters.

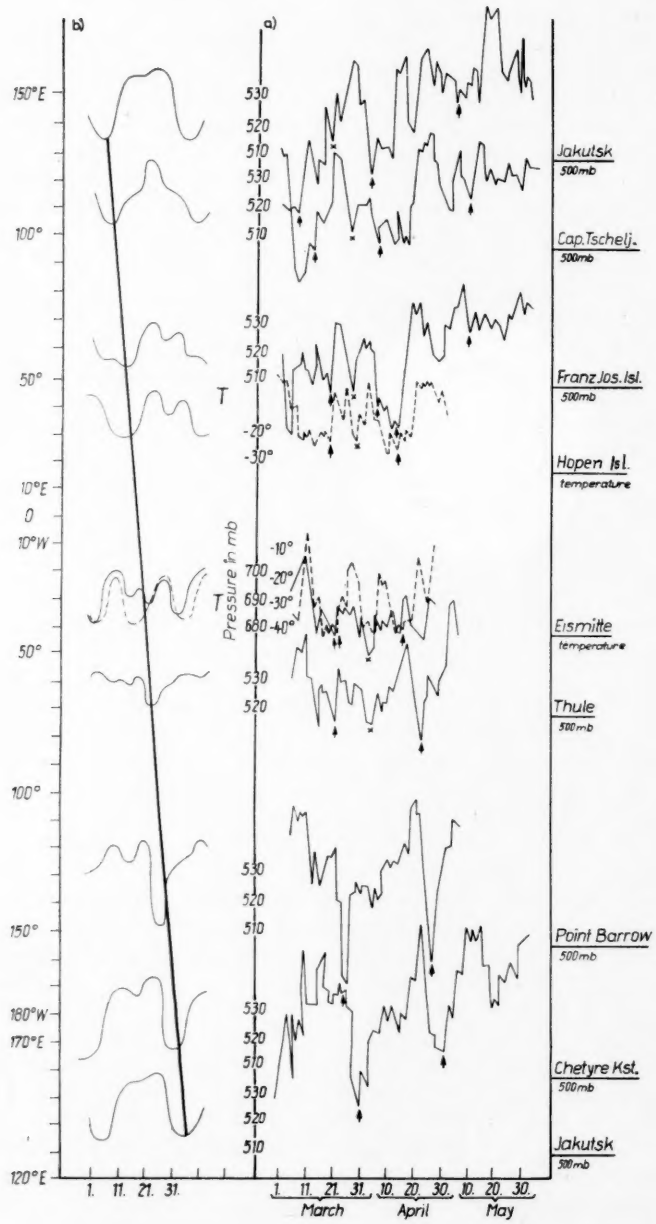


Fig. 19 (a). Height variation of the 500-mb resp. 700-mb level and temperature trend at some polar-stations

(b). Pressure and temperature variations averaged for a period of 30 days

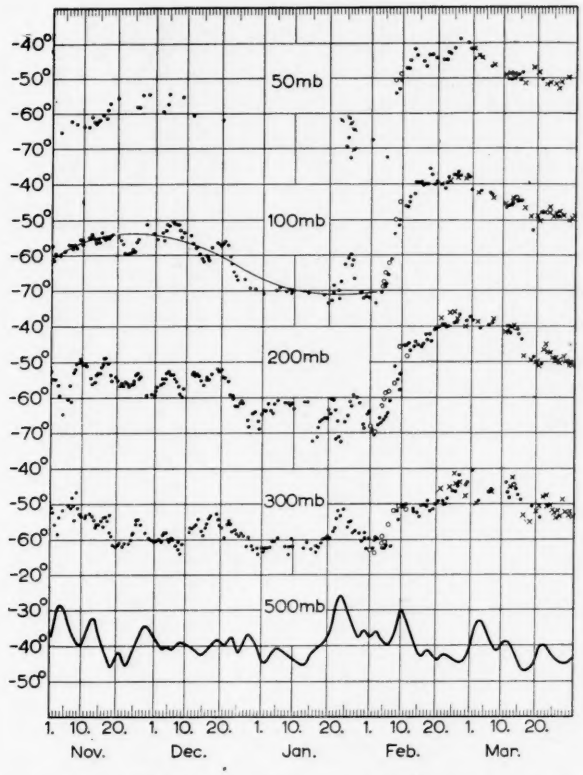


Fig. 20. Temperatures at some main isobaric surfaces over Alert, Ellesmere Land, 1 November 1951 to 31 March 1952

In *Fig. 22*, showing the height of the 50-mb surface on 3 February 1952, a cold stratospheric low center is situated near Thule and governs the stratospheric circulation over a great portion of the Arctic region. A week later (*Fig. 23*) the situation has nearly reversed. A stratospheric high-pressure center has replaced the cold low over northern Alaska, and on 21 February 1952 (*Fig. 24*) it reaches nearly the same position as did the cold stratospheric low three weeks before. The stratospheric low at the same time has been removed to northern Europe, and two days later the sudden stratospheric warming over Berlin began.[7] Within 7 days only, from 3 to 10 February, there occurred (*Fig. 25*) a height change of the 50-mb surface of more than 150 decameters in its center. It was not possible to explain the cause of this stratospheric warming and pressure rise in detail. It seems to be produced partly by advection, partly by subsiding processes in the stratosphere, but the warming occurring between 3–10 February was so pronounced that it may be that influences from ultraviolet radiation from the sun have participated too.

The other case which occurred in March 1954 developed in a similar way. This is shown by comparing *Fig. 26*, referring to the height of the 50-mb

Suppl.
Part. I
1956

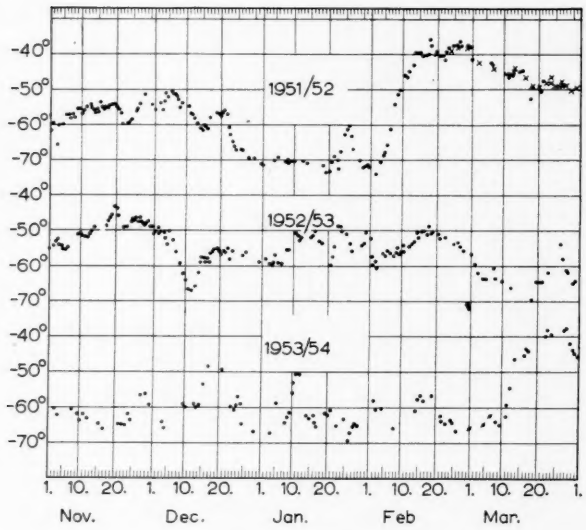


Fig. 21. Temperatures at the 100-mb level over Alert, Ellesmere Land, during winters 1951–52 to 1953–54

surface on 7 March 1954, with *Fig. 27*, representing the height of the 50-mb surface on 12 March 1954.

It may be that, to a great extent, these big stratospheric warmings are produced by a self-reinforcing process in that the temperature rise, which in the first moment may be small, will produce an associated pressure rise aloft. This pressure rise leads to the advection of warmer air moving along a geopotential line of higher value, and this will result again in a strengthening of the warm-air advection.

Probably these stratospheric changes will influence tropospheric weather phenomena, but this relation has not yet been solved clearly and will be a task of future research work.

Suppl.
Part. I
1956

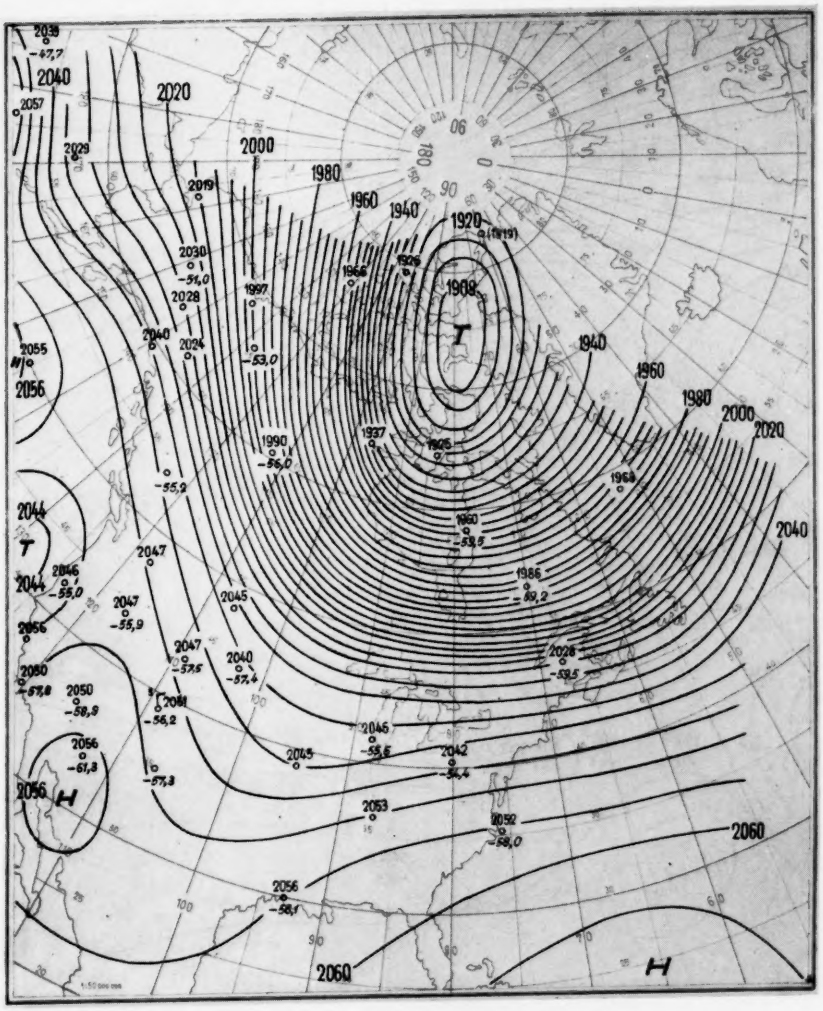


Fig. 22. 50-mb map, 3 February 1952, 0300 G.M.T.

(*Facing page* 116)

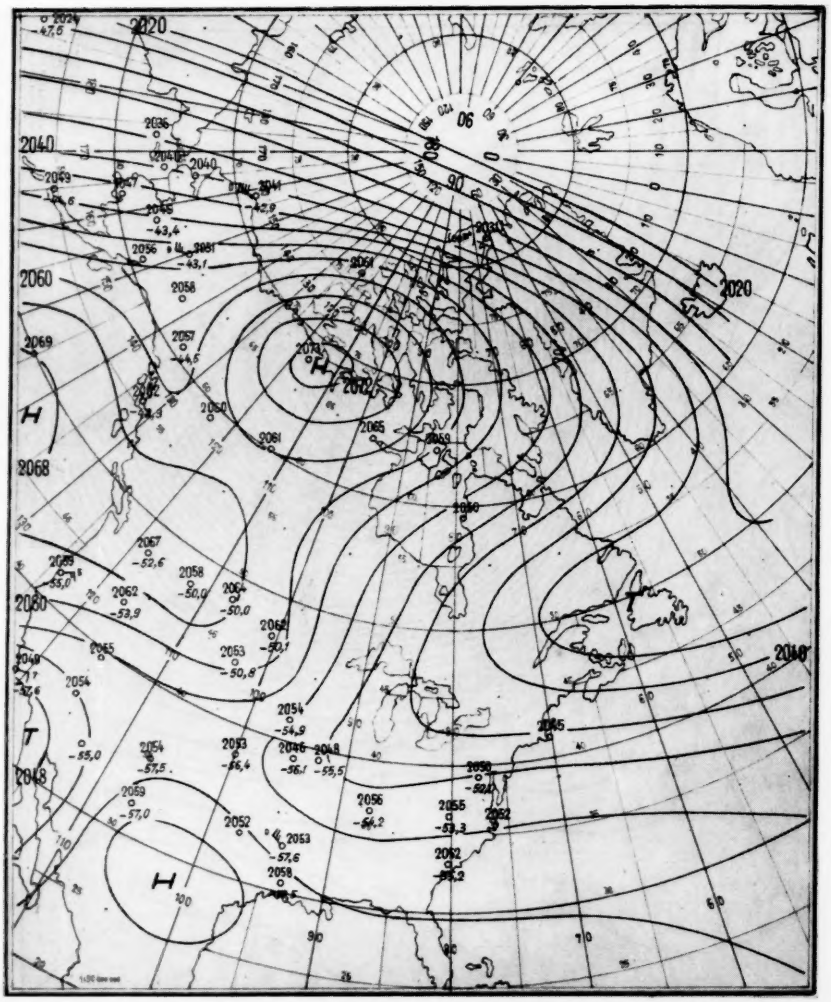


Fig. 23. 50-mb map, 10 February 1952, 0300 G.M.T.

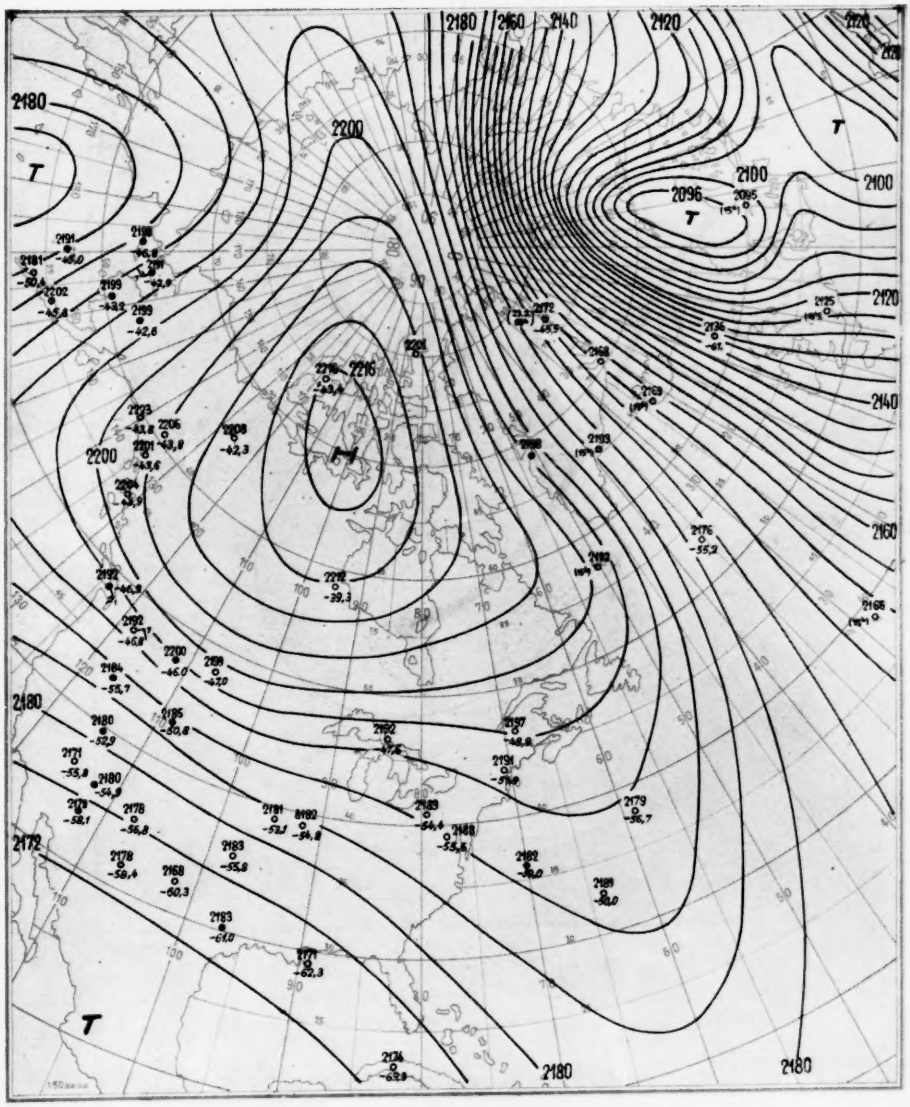


Fig. 24. 41-mb map, 21 February 1952, 0300 G.M.T.

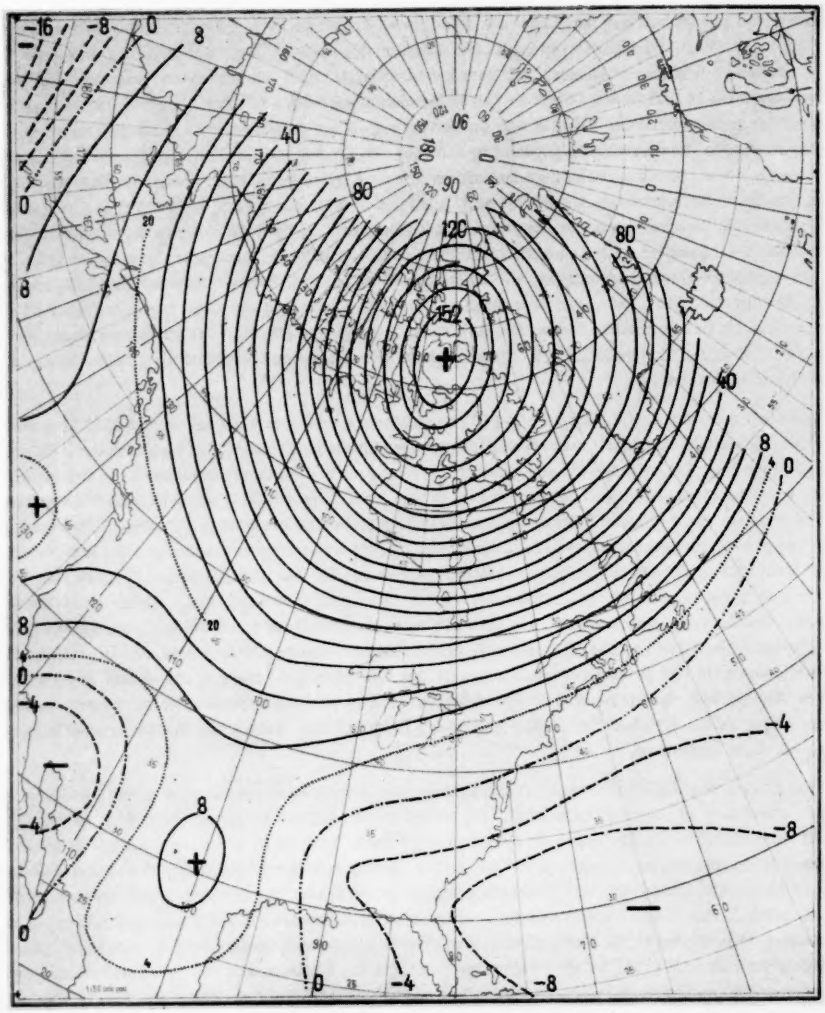


Fig. 25. Change of the height of the 50-mb surface, 3–10 February 1952 (gpdm)

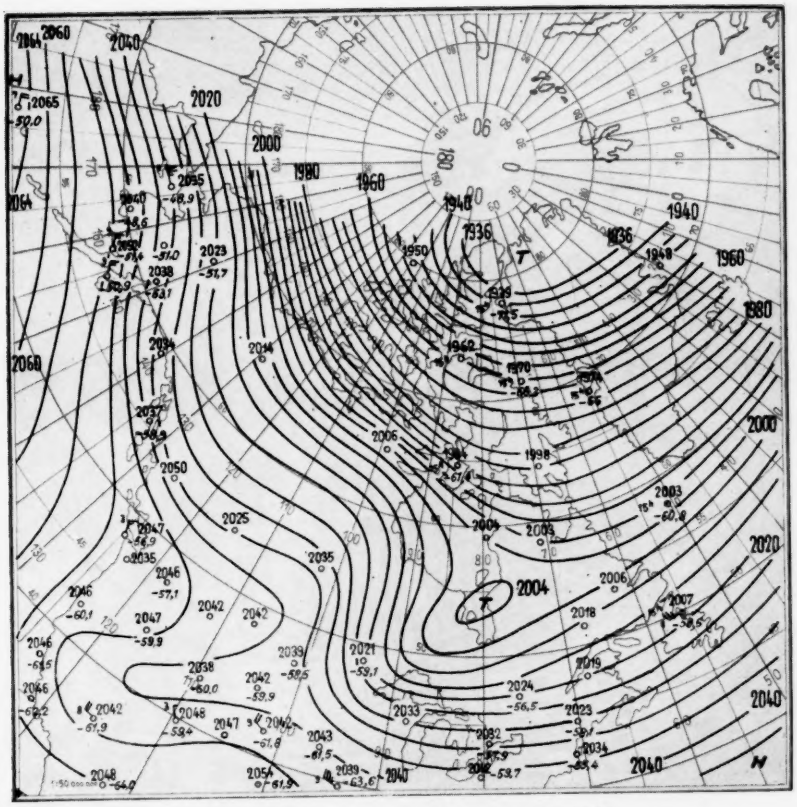


Fig. 26. 50-mb map, 7 March 1954, 0300 G.M.T.

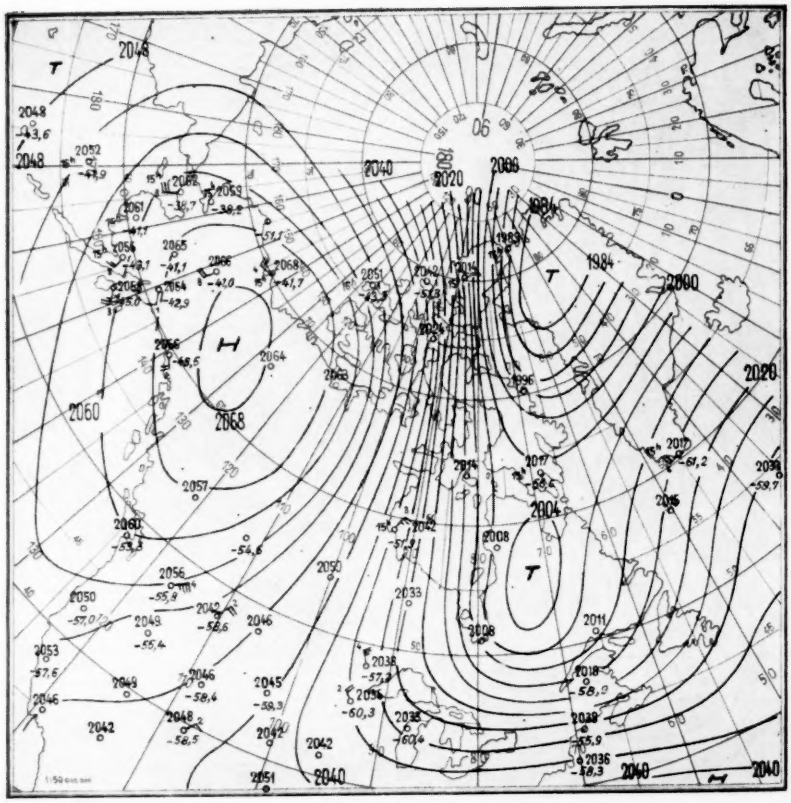


Fig. 27. 50-mb map, 12 March 1954, 0300 G.M.T.

(*Facing page* 117)

REFERENCES

[1] This map and the other monthly mean maps referred to are published by JACOBS, I. "5-bzw. 40-jährige Monatsmittel der absoluten Topographien der 1000-, 850-, 500- und 300-mb Flächen, sowie der relativen Topographien 500/1000 und 300/500 mb über der Nordhemisphäre und die Ursachen ihrer monatlichen Anderungen" *Met. Abh. des Inst. für Met. und Geophysik der Freien Universität Berlin* Bd. IV

[2] "Normalwerte des Luftdruckes auf der Nordhemisphäre für die Periode 1900–1939" *Met. Abh. des Inst. für Met. und Geophysik der Freien Universität Berlin* Bd. II/1

[3] U.S. DEPARTMENT OF COMMERCE, WEATHER BUREAU: *Normal Weather Charts for the Northern Hemisphere*, Technical Paper No. 21

[4] KLUG, W. "Der Mechanismus der Steuerung von hochreichenden Zyklonen" *Met. Abh. des Inst. für Met. und Geophysik der Freien Universität Berlin* Bd. III/2

[5] JORDAN, E. "An Observational Study of the Upper-wind Circulation around Tropical Storms" *J. Met.* Vol. V, p. 340, 1952

[6] WARNECKE, G. "Ein Beitrag zur Aerologie der arktischen Stratosphäre" *Met. Abh. des Inst. für Met. und Geophysik der Freien Universität Berlin* Bd. III/3

[7] SCHERHAG, R. "Die explosionsartigen Stratosphärenerwärmungen des Spätwinters 1951/52" *Ber. D. Wetterdienst in der US-Zone*, No. 38, S 51–63, 1952

[8] WEGE, K. Druck–, "Temperatur-und Strömungsverhältnissc in der Stratosphäre über der Nordhalbkugel" *Met. Abh. des inst. für Met. und Geophysik der Freien Universitäl Berlin* Bd. V/4

[9] U.S. DEPARTMENT OF COMMERCE, WEATHER BUREAU: Daily Series Synoptlc Weather Maps, Part 1: Northern Hemisphere Sea level and 500-mb Charts

Suppl. Part. I 1956

ARCTIC WEATHER-FORECASTING PROBLEMS OF THE AIR WEATHER SERVICE

ROBERT D. FLETCHER
Scientific Services, Air Weather Service, U.S. Air Force, Washington, D.C.

THE paper which I am presenting today is a recital of several problems which face the Air Weather Service of the U.S. Air Force in its attempts to predict the winds and weather of the Arctic. With respect to some of them which are obvious, I do not want to dwell on details but do wish to put them on record. Regarding a few of the other problems, attempts at their solutions have been made; as they come up in the discussion I should like to provide you with a few pieces of pertinent information which may be new to you.

There are many difficult problems of a logistical nature pertaining to direct or indirect support of Arctic forecasting operations. These include such items as the effort and expense involved in constructing observational installations, keeping them supplied, provision for power supply, arrangements for proper and reliable communications facilities and the like. The relative inaccessibility of the stations is the basic factor involved here. Another factor, which may vary from one weather service to another, is the abnormally short tour of duty of observing and forecasting personnel assigned to duty at remote Arctic locations. Familiarity with the terrain, climate and local weather aberrations has been adequately demonstrated as necessary for provision of the best weather service.

These logistic-type problems are the basic reasons for one of the most important problems facing the forecaster in the Arctic—inadequacy of weather data. Although the number of stations is increasing, the scattering is still too sparse for operational forecasting. In the Western Hemisphere north of the Arctic Circle there are now 17 upper-air stations. This means that there is one for about every 230,000 square miles in the Arctic as long as each one reports. To consider one of the middle-latitude areas, in the United States there are 70 upper-air stations, or one for about every 43,000 square miles, and there are eight surface for each upper-air station. Further, the distribution is fairly even in the United States, whereas it is decidedly uneven in the Arctic, where there are very large areas unrepresented. The ratio is more than 5:1 in favor of the U.S. distribution, which is estimated, by the way, to be about right for present numerical weather-prediction (NWP) work.

The "Ptarmigan" flights, from Fairbanks, Alaska, poleward and back, are helping considerably to ameliorate the data sparsity, especially with the current dropsonde observations. The tracks are being lengthened with the new B-50s, and we may be able soon to provide 300-mb data on the return

trip. We shall also have improved equipment aboard, including much better instrumentation for measuring flight-level winds.

The shortness of the period of record for many of the Arctic stations also is a handicap, especially with regard to formulating empirical methods of forecasting local weather phenomena. In addition, it precludes a knowledge concerning occurrences of unusual events. For instance, the "mean-annual" precipitation of Thule is 2·54 in., yet in May 1955, 10 in. of snow fell in one storm. The bay at Thule usually freezes over in September, but in 1955 did not freeze solid until late November.

Time delays in receipt of data, special data inaccuracies which may occur because of darkness and cold, and lack of certain types of equipment properly designed for Arctic use, all contribute to the general problem of data inadequacy.

What are some of the important forecasting problems for aviation? I shall briefly mention three of them. The first is middle-cloud forecasting—that is, for cloudiness between 10,000 and 20,000 ft which is important for flight operations in that layer. We are immediately handicapped by lack of humidity measurements at low temperatures. Again, even if the forecaster is fairly sure that clouds will be present he is not able to predict whether they will be dense, or in the form of the upper-level haze frequently encountered by aircraft in the Arctic regions. As yet we do not have adequate knowledge of typical air-to-air visibility in the haze or clouds.

Suppl.
Part. I
1956

We have made some attempts to associate middle-cloudiness with the vertical-motion charts turned out by the Joint NWP Unit in Washington. One trouble here is that the cloudiness forecast is difficult to verify. Further, the detail in the analysis is not nearly sufficient. So far, we have not had much luck with this method. Perhaps the best and most practical we have found is a semi-empirical one, due to de Bellefeuille, in which marked geostrophic advection of vorticity at 500 mb is associated with the presence of clouds, and vice versa. When the advection is ill-defined, one finds that a forecast of "no change" in 24 hr is usually pretty good.

As aircraft start to operate at higher altitudes, we must begin to collect information on the vertical extent of haze and cirrus, frequency of occurrence of the phenomena, and visibility expectancies in them.

The second problem is one of the most critical which faces any service concerned with forecasting for aircraft, especially of the jet variety—local terminal prediction of ceilings, visibilities and strong surface winds. At Thule, for instance, one of the really difficult problems is the forecasting—even a short period ahead—of the occasional hurricane-force winds which blow across the approach zone of the runway and which begin and end suddenly, in a matter of minutes. Winds of 50–80 knots are often localized within distances of less than 10 miles. The strongest winds blow at an angle of 70 or 80° to the runway *across* the mountains which are parallel to the runway, rather than along the valley. Sometimes they blow at elevations of 1000–2000 ft and occur at the base only several hours later, or sometimes not at all. No change in magnitude of the pressure gradient is apparent. It is very difficult to explain the direction and violence of these gravity winds.

At this and several other stations summer fog sometimes moves in rapidly and may be very dense. It may occur as a shallow, dense bank or may be

associated with rain and thick cloud layers. Its behavior is exceedingly erratic—occasionally it persists for several days in a stretch. It occurs with (usually) recognizable types of local circulation, but these are difficult to predict because of few data and lack of adequate prognostic techniques. There is a certain amount of diurnal variation in weather here; summer fog has about 30 per cent greater incidence during the "night" hours of 1800–0600, but the variation is not dependable enough for forecasting use.

A similar problem frequently occurs in the Alaskan valleys. Mitchell, for example, has discussed the extremely localized, strong winds which blow along a long narrow path, in a shallow layer, near Fairbanks. Again, gales will be experienced in the Matanuska Valley near Anchorage; a mile away nearly calm conditions will prevail. Gales will blow out of Turnagain Arm, just south of Elmendorf Field at Anchorage where light variable winds are being reported.

For forecasting ice-fog, and for predicting maximum and minimum temperatures which are important for the winterizing of engines, no really good methods are known. At Alaskan stations, differences of up to 22° in minimum temperatures have been observed at sites 2000 ft apart. The all-important surface winds are essentially unpredictable because of the difficulty in forecasting katabatic or even gradient winds.

The influence of local terrain on precipitation, by the way, can be extremely great. For example, Elmendorf Field has an average of 15 in. of precipitation per year. Thirty miles away, separated from Elmendorf by the Chugach Range with peaks averaging only 7000 ft, lies Whittier which has an average annual amount of 150 in. In 1954–55, Elmendorf had its greatest snowfall on record: 132 in. Whitter's report was 832 in.

I do not want to belabor the point with more examples. But the local variations are really extreme, and many of the weather occurrences are of the greatest significance in aircraft operations in and out of airfields. In the cases of a few airfields, meteorologists have developed useful statistical relationships between observational data and operationally important parameters of a few hours later; usually, however, the data-sparseness problem precludes this approach. Subjective association of the operationally important factors with predicted synoptic patterns through recourse to either experience or meteorological theory is the more frequent approach. Here, too, insufficiency of data is a very real obstacle as is our relative unfamiliarity with the synoptic features of the Arctic. There is great need for development of methods of forecasting the local phenomena, of providing the data required for applying the techniques which are developed, and in spreading information on successful techniques to all concerned.

The third very important problem is the prediction of upper-air winds. In this connection we ran, by the way, an evaluation of NWP over the Arctic to see how it compared with the subjective product of a large weather central in the U.S. The prognoses were for 36 hr in June 1956. The two prognoses were about equal for 1000 mb, but the NWP product was noticeably better at 500 mb.

Understanding of large-scale processes would certainly aid the overall problem of forecasting. Thus, we should like to know more about the structure of Kochanski's high-level cold vortex of winter, and how it varies. Too,

how can we tell when the unusually warm temperatures sometimes observed at Arctic stations will occur, and what bearing do they have on later occurrences at lower levels? A related problem is the development of improved methods of deducing (perhaps by climatological analysis) high-altitude pressures, temperatures and winds from meteorological soundings which have terminated at low levels.

In an attempt to learn more about the structure of very high-level winds, Kochanski recently studied two series of wind and temperature soundings made with special equipment; the layer considered was from 700 to 5 mb. The data were from a line of four stations: Patrick Air Force Base, Florida; Belmar, New Jersey; Goose Bay, Labrador; and Narsarssuak, Greenland. The two month-long series were for August 1954 and January 1955. A summary of some of the more important information derived from the investigation follows. The summary includes comments with regard to the middle- and low-latitude stations in order that the high-latitude data can be viewed in the perspective of the zonal patterns of the Northern Hemisphere as a whole.

(1) In winter Patrick was characterized by patchy easterlies in the layer from 60,000 to 90,000 ft; this layer was topped by strong westerlies.

(2) At Belmar the winter easterlies were more extensive in depth, especially during the first half of the month.

(3) At Goose Bay, in January, there were still more easterlies and the winds were weaker.

(4) At Narsarssuak, north-westerlies predominated above 80,000 ft.

(5) In summer Patrick's winds were all from the east and were remarkedly steady in both speed and direction above 80,000 ft.

(6) At Belmar, in summer, the winds changed from west to east above 70,000 ft and the wind strengths were weaker.

(7) At Goose Bay, in summer, the winds again changed from westerlies to easterlies at around 70,000 ft but their strengths were much weaker and there was somewhat more variability in direction at all levels for which data were available.

(8) At Narsarssuak, in summer, winds were mostly light and variable although there was a tendency for easterly components to dominate above 60,000 ft.

(9) At all stations the 24-hr vector wind changes were relatively small (less than 20 knots) between 50,000 and 110,000 ft, both summer and winter.

(10) At all stations, both seasons, small 24-hr vector changes corresponded to low scalar wind speeds except those in summer at Patrick and Belmar, where the stability of the easterlies at heights above 50 mb was very marked.

(11) For all the stations grouped together, the scalar wind speeds reached a maximum near the tropopause level, and a minimum near 25 mb in winter and near 50 mb in summer. All 24-hr changes reached a maximum near 300 mb and decreased rapidly thence upward.

The preceding material illustrates some of the recent work done to provide us with background material necessary for the problem of prediction. At the same time, it gives us an estimate of the value of persistence as a forecasting tool; as far as wind magnitudes are concerned, it demonstrates that persistence indeed can be rather successful at higher levels.

Suppl.
Part. I
1956

The value of charts of departures-from-normal (DNs) of heights of a pressure surface has recently been demonstrated by D. E. Martin in his attempts to relate DNs at "key" locations with those at other latitudes and longitudes in the Northern Hemisphere, including such areas in Arctic regions. In his work Martin utilized 700-mb data consisting of 5-day means at 10° latitude and longitude intersections. He has pointed out that extensions to other pressure surfaces and use of space- instead of time-mean data also appear feasible. In its application to forecasting problems, the method depends on successful forecasts of DNs at the key locations (intersections of certain meridians with the 40°, 50° or 60° parallels), whence Martin's charts enable the forecaster to state the expectancies of the DNs at locations for which predictions are desired. Combination of the expected DN pattern with the "normal" pattern provides the forecaster with a basis for prediction of wind fields as well as, of course, a basis for a subjective estimate of future weather.

One set of Martin's charts is in the form of a series of maps containing isopleths of the per cent occurrence of greater than 200-ft height anomalies, one chart for each case of either above—or below—normal (in excess of 200 ft) heights at each key location. Examples of conclusions drawn from the set of charts are:

(1) Between May and September, there is a 70–80 per cent probability of occurrence of above-normal 700-mb 5-day-mean heights over Greenland when, at 50°N latitude, negative departures are observed between either 150° and 160°W longitude or 80° and 90°E longitude.

(2) Between November and March, there is a 70–80 per cent chance of negative anomalies over the Arctic Ocean when, at 50°N, heights are above normal between 140° and 150°E.

(3) For the same winter months, there is a 70–80 per cent chance of above-normal departures over the Arctic when, at 60°N, heights are above normal between 30° and 40°W.

(4) For the same months, there is nearly a 90 per cent probability of negative anomalies over the Arctic when, at 60°N, negative anomalies are observed between 40° and 50°E.

Martin has prepared another set of charts from the same data, relating the average magnitudes of height anomalies, in the form of isopleths over the Northern Hemisphere, to anomalies observed at the same key locations. From these charts conclusions such as the following can be drawn:

(1) From December to February, a weak westerly 700-mb circulation in Arctic regions can be expected, on the average, when positive anomalies are observed at 40°N between 160° and 170°E.

(2) Similarly, and for the same months, strong southerly flow from the western Pacific into the Arctic is associated with positive anomalies at 40°N between 0° and 10°E.

(3) Similarly, and for the same months, strong southerly flow over Greenland is associated with negative anomalies at 50°N between 90° and 100°W.

Further extension of Martin's statistical work to include more key areas, more pressure surfaces, use of basic data other than 5-day-mean heights

and introduction of time lags, should make available to the forecaster even more useful prediction tools.

I have tried to give you a picture of some of the most important Arctic forecasting problems which face us in the Air Weather Service, realizing that they must be important, too, to many of you. To the numerous valuable results of research presented here at this symposium I have also added the results of several other efforts to give us answers to these problems. The stimuli and spreading of information resulting from symposiums such as this one will, I am sure, lead to accelerated work on these and other important Arctic meteorological problems and a more rapid attainment of useful results than would be the case if each of us proceeds along his own uncoordinated path.

Suppl.
Part. I
1956

ARCTIC WEATHER ANALYSIS

RICHARD J. REED
University of Washington, Seattle, U.S.A.

A survey of the literature on the synoptic meteorology of the Arctic and inspection of the synoptic charts of several meteorological organizations reveals that no special methods of weather analysis are used in or recommended for polar regions. Standard surface analysis consists of isobaric and frontal analysis and standard upper-air analysis of contour analysis only. Canadian analysts also prepare frontal contour charts on a regular basis and use these to locate three basic fronts on both surface and upper-level maps. One of these fronts, the continental Arctic front, separates continental Arctic air from maritime Arctic air.

A notable feature of the polar troposphere is the low-level inversion which blankets all parts of the Arctic in winter and the Arctic Sea in summer. Although the inversion has attracted considerable attention in Arctic research, it generally is not represented on synoptic charts.

Much of the research by members of the Arctic Meteorology–Climatology Project at the University of Washington has been concerned directly or indirectly with problems of Arctic analysis. Numerous vertical cross-sections have been prepared in an attempt to clarify the baroclinic structure of polar weather systems. The results of 63 cross-sections constructed along a line from Alert, Ellesmere Island, to Kodiak, Alaska, during the period February–June 1955 are summarized and some sample cross-sections shown It is concluded that the baroclinic structure is complex and usually not well represented by conventional frontal models. The view is presented that this disagreement is not a peculiarity of Arctic weather systems but is characteristic of middle-latitude systems as well.

The question of the adequacy of the polar observational network for numerical weather prediction is discussed. Results of 18 barotropic forecasts by the Fjørtoft method indicate that sufficient data is available over the Northern American half of the Arctic for numerical methods to be employed successfully. For the 18 cases the highest correlation between 24-hr observed and predicted 500-mb height changes is 0·82 and the lowest 0·20 Cases of poor verification seem attributable to sudden developments in the pressure field rather than to uncertainties in specifying the initial contour patterns.

IT is apparent from an examination of the literature on the synoptic meteorology of the Arctic and from an inspection of the synoptic charts of a number of meteorological organizations that no special techniques are employed in the analysis of Arctic weather data. The atmosphere knows no impermeable boundary which separates Arctic air-masses from those which lie to the south. On the contrary, a perpetual large-scale exchange of air occurs between polar and adjacent regions as Arctic air periodically surges southward and extra-polar air streams northward to take its place. Thus from the synoptic viewpoint there is justification for regarding the subject of Arctic analysis as merely an extension of the subject of analysis in general, and such appears to be the prevailing attitude.

Sup
Par
195

Standard surface analysis, as revealed by a survey of the charts of several meteorological services, consists of isobaric and frontal analysis. Warm, cold and occluded fronts are the basic ones entered on the weather maps, usually more or less in conformity with the classical Norwegian cyclone model. On some charts a distinction is made between upper fronts and surface fronts. Canadian analysts also enter a feature known as a trowal which represents the projection on to the surface map of the locus of a warm air crest.

At upper levels it is customary for only contour-heights to be represented on published maps. In some instances, isotherms are drawn as well. Although upper-level fronts are occasionally projected on to surface maps, it is not usual to analyze fronts on the upper-level charts themselves. A noteworthy exception to this statement is the practice of Canadian analysts who, with the help of frontal contour charts, delineate three fronts on all tropospheric charts. These fronts are analyzed as continuous in both space and time although GODSON,[4] a leading spokesman for the group, states that the fronts are regarded as continually forming and dissolving.

From the foregoing brief discussion it is apparent that some differences in methods of analysis exist but that in general surface analysis adheres to the Norwegian system, while at upper levels emphasis is on the flow pattern with lesser attention given to fronts.

Suppl.
Part.I
1956

Having established current analysis procedures, we may ask how well these procedures perform the functions for which they are intended and how they may possibly be improved. All meteorologists would agree that the main purpose of analysis is to represent, in a concise, graphical manner, those features or elements of the atmosphere which are regarded as important in determining its future state. However, it is questionable whether there is the same unanimity of feeling regarding what constitutes a satisfactory system of analysis.

What the analyst enters on his map and the significance he attaches to it depends on his viewpoint concerning atmospheric processes, and since the past 15 years have witnessed a radical change in this viewpoint, at least in many quarters, it is safe to say that a considerable spectrum of opinion now exists as to how maps are best analyzed. Before the new outlook developed, the synoptic meteorologist thought almost entirely in terms of certain gestalt concepts, foremost among these being the ideas of fronts and air-masses. It was believed that the atmosphere was composed of large masses of air with nearly homogeneous thermal and moisture properties. Between these masses of air, or air-masses, were narrow zones of transition or fronts, and it was supposedly the shearing instability at fronts which gave birth to storms. The principal front was that which separated tropical and polar air-masses, and this was named the polar front. Clouds and hydrometeors were related to the vertical motions at fronts and to the hydrostatic stability within the air-masses.

Such in brief outline was the classical viewpoint which enjoyed such popularity in the 1920s and 1930s. However, as upper-air observations became more plentiful the phenomenon of upper waves in the westerlies was recognized and soon claimed an increasing share of the meteorologist's attention. In the early 1940s a new era dawned when ROSSBY[6] successfully applied vorticity concepts to the problem of long barotropic waves. There

followed next the important contributions of SUTCLIFFE[9] on the dynamics of baroclinic waves and the exciting development of numerical weather prediction by CHARNEY[2] and others.

Fronts have played little part in these recent advances; however, not because of a desire on the part of investigators to ignore them, but simply because they are awkward to handle in dynamic models. The degree of success achieved by these models makes it clear that fronts do not have the significance formerly attached to them. This does not mean that fronts do not exist or that they are not important in certain problems. It does mean, though, that a new outlook is required concerning the role of fronts in the large-scale atmospheric developments. In this new outlook relatively broad and diffuse baroclinic zones are regarded as sufficient for cyclone development. It is recognized that in many instances a pre-existing front will be the preferred location for the appearance of the incipient cyclone. However, it is also recognized that on other occasions the front will develop simultaneously with the cyclone due to the deformation of the wind field in the region of the growing disturbance. As regards analysis, the result of the recent advances is to shift emphasis away from delineation of synoptic entities, such as fronts and air-masses, to the representation of continuous fields of the principal atmospheric variables, mainly temperature and pressure (or contour-height).

The speaker feels that the modern viewpoint, as just presented, will inevitably supplant the classical outlook. If this is the case, the basic need in Arctic analysis, as in middle-latitude analysis, is for accurate specification of temperature and contour fields. Fortunately, within the limitations imposed by the data, it is possible to analyze these fields with a considerable degree of accuracy, certainly with greater confidence than is the case with fronts.

To illustrate the data requirements of a numerical prediction model and the special problems that arise in the Arctic in connection with these requirements, let us consider the model devised by SAWYER and BUSHBY.[8] The basic data required by this model are 1000-mb and 500-mb heights and the thickness of the layer 1000–500 mb. Obviously only two of these quantities are independent, and once these are read from analyzed charts the third may be obtained by addition or subtraction. Which two are chosen as primary variables will depend on the amount and type of data available in the area under consideration.

In areas where radiosonde observations are plentiful it probably makes little difference in the final result which two are taken as independent; and since it is simplest to analyze the 1000-mb and 500-mb charts and to obtain the thickness by subtraction, this method is to be recommended. In the Arctic, however, the situation regarding data is not the same, and alternative procedures must be considered. Over most areas there are more surface observations than upper-air and the station density is thin by middle-latitude standards. Under these circumstances differential analysis working up from the surface chart should give the optimum results, especially when hodographs are available to aid in orienting and spacing the thickness lines.

On the other hand, over a large portion of the Arctic Sea, air reconnaissance by the "Ptarmigan" flight is the principal source of weather information. The flight route is from the north coast of Alaska to 85°N and return,

one leg being flown at the 500-mb level and the other at the 700-mb. Three dropsondes are made on each leg, and frequent wind measurements are taken as well. As many as 20 wind observations may be received from one flight.

In these circumstances it is obvious that the 500-mb chart may be analyzed with greatest confidence, and thus, if differential analysis is employed, this must replace the 1000-mb chart as the starting chart. It might also be pointed out that from the standpoint of the Sawyer-Bushby model, it would be preferable if one leg of the "Ptarmigan" flight were carried out at a lower level than 700 mb; for instance, 900 mb or below. This would allow the basic data of the model to be specified with greater accuracy. In any case it can be seen that the main analysis requirements of a numerical prediction model can be met by techniques that are already familiar to the analyst.

The opinion has been expressed that the modern viewpoint places greater emphasis on continuous fields of the fundamental meteorological variables and less on gestalt concepts such as fronts and air-masses. However, as long as fronts can be defined in an objective manner, for instance by considering them to be first-order discontinuities in temperature, then there is justification for entering them on maps. Moreover, it is probably true that most forecasters would be reluctant to omit fronts from their maps. Even those who consider fronts to be only a detail of the baroclinic structure realize that in practical forecasting details are often of critical importance. The prediction of large-scale developments is only one part of the forecast problem.

Suppl.
Part. I
1956

Granting then that fronts are still a factor which must be reckoned with in weather analysis and forecasting, I would like next to ask the following questions: Are surface fronts currently being analyzed in a satisfactory manner? Do the conventional three-dimensional models of fronts present a true picture of the usual baroclinic structure of the atmosphere? And how much information does the front actually give concerning the weather? In answer to the question of how well fronts are analyzed, we may compare, as has often been done, analyses of the same weather situation made independently by several competent analysts. The analyses selected for comparison were prepared by three national weather organizations. The method of comparison has been to select three areas of similar size and shape, one over the Arctic Sea where data are scarce, a second over the Canadian Archipelago where the station network is relatively dense, and a third over north-western Europe where observations are plentiful. Sheets of paper with the test areas printed on them were prepared for each day of two different months. The fronts within each area, as analyzed by the three organizations, were transcribed on to the sheets, different colours being used for the analyses of each service.

The original intent was to compare the analyses front by front, measuring average deviations in position. However, the differences in analyses were so great that this course was not feasible, and instead the results of the tests have been summarized in the following simple fashion: A front is considered to be the same front on two different analyses if a front of any description and in any orientation or a frontal complex is located in the same general area. Next the total number of fronts or frontal complexes for each month

have been counted. Finally, the totals have been divided into the number of cases in which only one service analyzed a particular front, in which two services analyzed it, and in which all three analyzed it. The results for the first month, February 1956, reveal a total of 27 fronts in the Arctic Sea sector. In 11 cases only one analysis contained the front, in 8 cases only two and in 8 cases all three. This result is not surprising when the sparsity of data is considered. However, in the second area, the Canadian Archipelago, where data is more abundant, the results are much the same except that fewer fronts were analyzed during this particular month—a total of 15 fronts with 10 analyzed by only one service, 3 by two services and only 2 by all three. It might be anticipated that the European sector would show distinctly better agreement among the various analyses, but such is not the case. The figures here are a total of 43 cases with 18 fronts entered by only one service, 19 by two and only 6 by all three.

February 1956 was characterized by an extremely anomalous circulation pattern. The European continent experienced bitter cold, while unusually warm and stormy weather prevailed near the North Pole. The second month tested, April 1956, was more normal, especially about the Pole where a typical anticyclonic régime was found in place of the cyclonic pattern of February. The figures for this month show far fewer fronts in the Arctic Sea sector and also better agreement in the European sector, although in the latter area the results are hardly gratifying if the lenient interpretation of what constitutes similar fronts is taken into account. *Table 1* summarizes the results:

Table 1. Comparisons of the Frontal Analyses of Three National Weather Organizations

February 1956

	One Analysis	*Two Analyses*	*Three Analyses*	*Total*
Arctic Sea	11	8	8	27
Canadian Archipelago	10	3	2	15
North-western Europe	18	19	6	43

April 1956

	One Analysis	*Two Analyses*	*Three Analyses*	*Total*
Arctic Sea	5	1	0	6
Canadian Archipelago	14	8	5	27
North-western Europe	10	9	18	37

Next comes the question of the adequacy of conventional, three-dimensional frontal models. The speaker will take as a definition of a front, one which has been suggested by BOVILLE.[1] Thus a front is defined as:

(1) A three-dimensional hyperbaroclinic zone with a first-order discontinuity in the temperature and wind fields.

(2) A quasi-substantial surface which moves with the wind.

(3) A reasonably continuous feature of the chart both in space and time.

By a *conventional*, three-dimensional front, I will mean a zone of the type described above which slopes upward towards the cold air and which starts at the earth's surface and passes through a substantial depth of the troposphere.

My contention is that zones which meet all these requirements are a rarity. They do not represent the usual baroclinic structure of the atmosphere, but greatly oversimplify it. Carefully constructed vertical cross-sections are the best means for checking the correctness of this assertion. Since these are prepared mostly for research studies, the contact which any one meteorologist can have with such sections is rather limited, and therefore my belief may be somewhat biased. However, in support of my claim, at least as far as the Arctic is concerned, are the results of 63 cross-sections, extending from Kodiak, Alaska, to Alert, Ellesmere Island, which have been prepared by the staff of the Arctic Meteorology-Climatology Group at the University of Washington during the period February–June 1955.

It is difficult to summarize the features of these sections in an objective manner so that the stated results are admittedly somewhat arbitrary. Of the 63 sections, six contained frontal zones in the area of the cross-section, though on all six the lower part of the front was over the Gulf of Alaska beyond the edge of the section. Quite possibly the number would be reduced

Suppl.
Part. I
1956

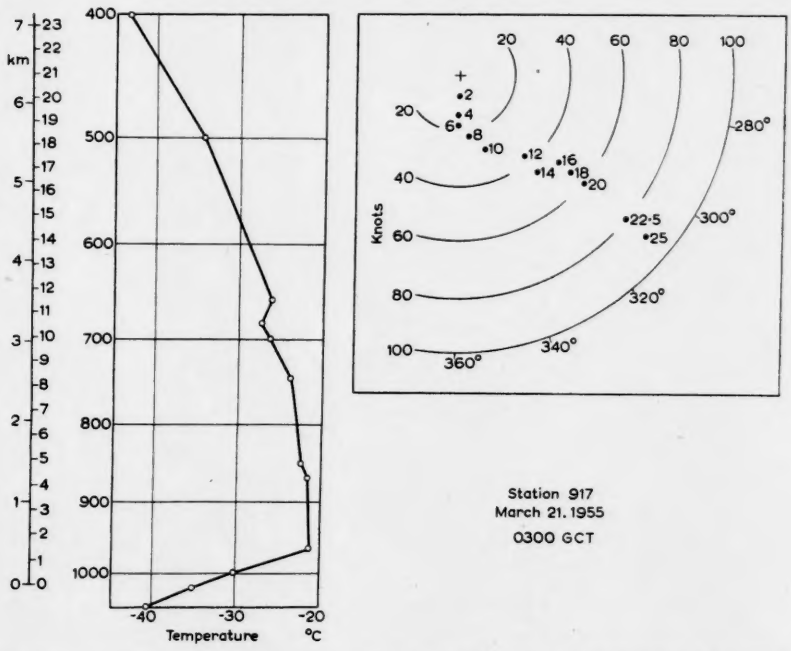


Fig. 1. Temperature sounding and hodograph at Eureka N.W.T. (917) at 0300 G.C.T., 21 March 1955. Heights in thousands of feet are given next to the points on the hodograph

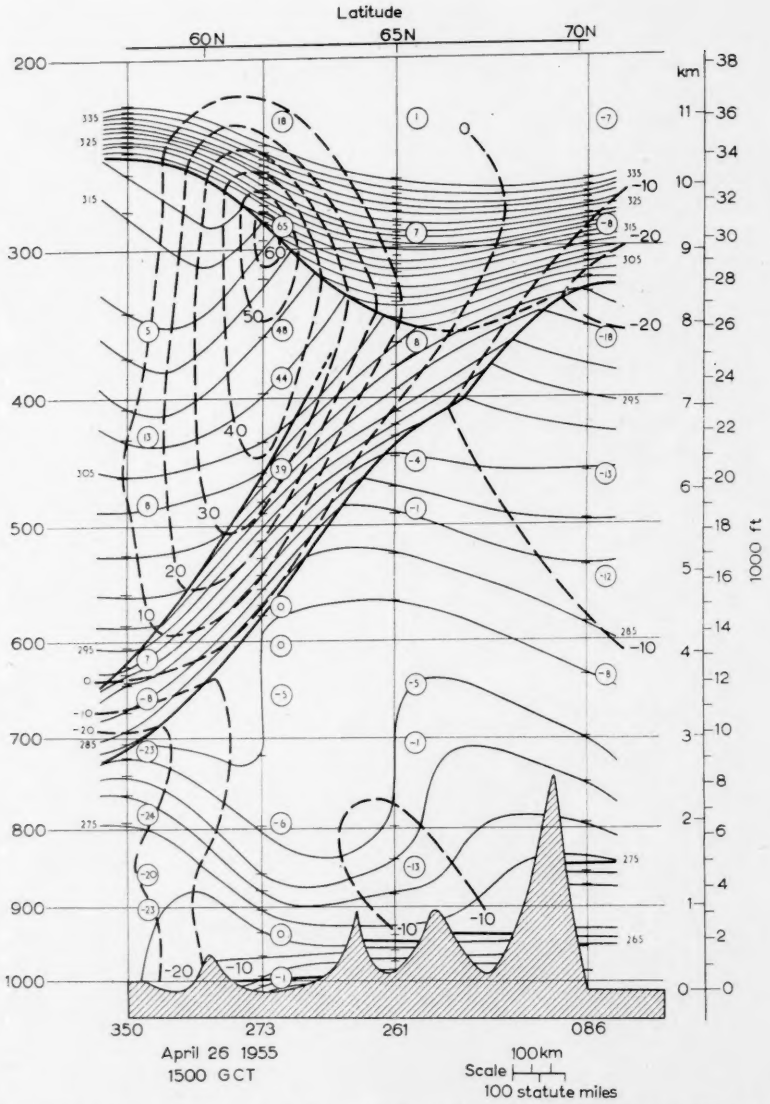


Fig. 2. Vertical cross-section from Kodiak Island, Alaska (350), to Barter Island, Alaska (086). Thin solid lines are isopleths of potential temperature in °A. Medium dashed lines are isotachs of wind component normal to the section in knots. Heavy solid lines are tropopauses, frontal boundaries or inversion surfaces. Weak discontinuities are denoted by heavy broken lines. Numbers in circles are normal wind components in knots

if data for the ocean area were available. Of the remaining sections eight showed well-developed baroclinic zones without sharp boundaries; the other 49 had weak or diffuse baroclinity. The hodograph in *Fig. 1* is presented as an example of pronounced baroclinity without true frontal structure. Such hodographs are not uncommon in the vicinity of polar vortices.

Fig. 2 illustrates a case of true frontal development and shows the method of cross-section analysis used at the University of Washington. The attempt is made to be as objective as possible. Along each vertical potential temperatures, as read from the appropriate sounding, are marked at 2°C intervals. Actual wind components normal to the section are plotted beside the verticals. Isentropes are drawn first in the vicinity of the verticals, use being made of the thermal wind relationship to estimate the slopes of the various isentropic surfaces. Next the line segments are joined in the areas between stations, care being taken to preserve the baroclinic features suggested by the disconnected lines. Finally, a wind analysis is drawn which is consistent with the temperature distribution.

A frontal zone of about 100 mb in depth is plainly discernible on the diagram. It is interesting to note in passing that the tropopause is extremely sharp in the core of the jet, a feature that is not uncommon despite the emphasis in synoptic literature on fractured and multiple tropopause structure near the jet stream. Also, not an inconsiderable portion of the vertical wind shear connected with the jet stream is contained in a lesser baroclinic zone just below the tropopause. This too is a common feature of cross-sections analyzed in the above fashion.

Two examples of sections exhibiting more complex baroclinic structures will now be considered. These were not constructed initially to emphasize the complexity of the situations. Rather they were chosen because the surface maps showed unusually well-defined cold fronts, and it was desired to study their three dimensional structures.

The first case, which occurred on 18 August 1955, is shown in *Fig. 3.* Wind hodographs were available at four of the five stations which lie on a line from Point Barrow, Alaska, to Fort Smith, Canada. At the fifth station geostrophic winds were added from constant-pressure charts. The cold front, which moved from left to right across the section, appears to resemble the Canadian trowal more than the text-book cold front. The leading edge of the cold air arrives first aloft. There is no evidence of a concentration of temperature in a narrow zone sloping westward toward the dome of cold air.

The second case occurred in the same region on 16 January 1956. Again the front travelled from left to right and again the cooling began first at upper levels. The baroclinity is extremely complex. Part of the complication is due to the presence of a superadiabatic layer at Barter Island. Since superadiabatic lapse rates are encountered quite frequently on soundings in the Alaskan area, this feature is believed to be real.

The presence of low-level inversions on the two sections and of a moderately high mountain range parallel to the line of the sections at the left end perhaps contributes to some of the unusual aspects of the diagrams. However, the speaker doubts whether orographic features or surface friction are the only factors involved in the initiation of the cooling at upper levels. Data from such low-lying areas as the Gulf of Alaska, eastern United States

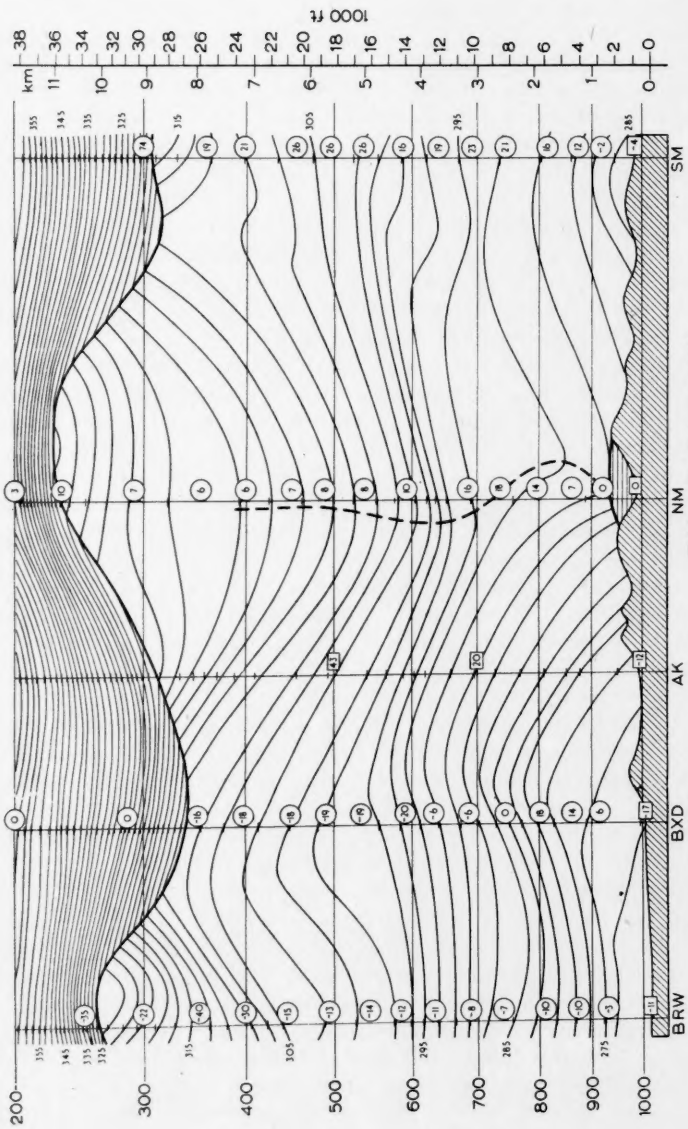


Fig. 3. Vertical cross-section from Barrow, Alaska (BRW), to Fort Smith, Canada (SM). Thin solidlines are isopleths of potential temperature in °A. Heavy solid lines denote tropopause and inversion surface. Heavy dashed line represents leading edge of area of cold advection. Actual wind components normal to the section are shown in circles; geostrophic components in rectangles

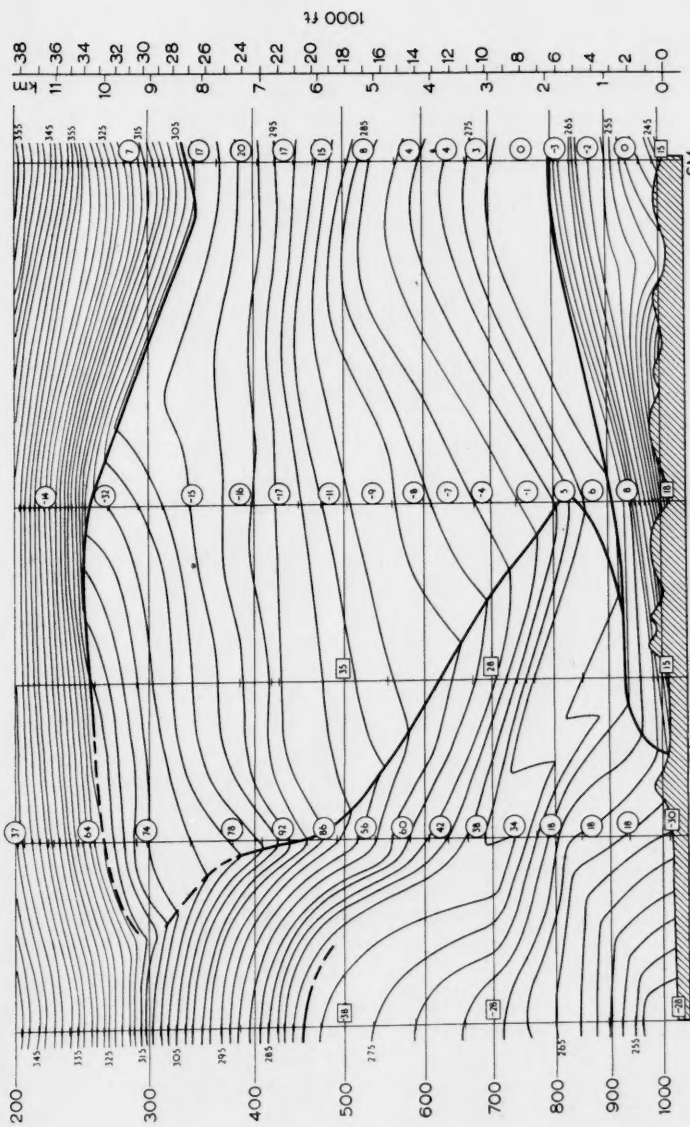


Fig. 4. Vertical cross-section from Barrow, Alaska (BRW), to Fort Smith, Canada (SM). Thin solid lines are isopleths of potential temperature in °A. Heavy solid lines denote tropopauses, frontal boundary and inversion surfaces. Weak discontinuities are represented by heavy broken lines. Actual wind components normal to the section are shown in circles; geostrophic components in rectangles

and northern Germany have revealed this phenomenon, and sometimes the initial cooling is at 10,000 ft or above. Also it is doubtful if the phenomenon can be reconciled with the classical cyclone model by regarding it as the upper cold front of a warm type occlusion since in many cases there is no evidence of the occlusion process. PENNER[5] has stated that "cases of true occluded fronts as in the classical model are extremely rare over North America and adjacent regions". German meteorologists have long recognized the phenomenon under discussion and describe it as "voreilen kaltluft".

Next, as regards the question of how much information frontal analysis gives concerning the weather, we may cite the results of the British Meteorological Research Flight 1950–52. One of the principal conclusions reached by SAWYER[7] on the basis of the flight data is that the "cloud structure of fronts is very variable". In fact the upper portions of fronts were found to be characteristically dry rather than moist, a finding which is in accord with the speaker's own researches on frontal zones of North America. It may be pointed out too that radar studies have revealed the complexity of the relationship between fronts and precipitation patterns.

To sum up, then, the opinion is expressed that the analysis of surface fronts is not in a completely satisfactory state, that idealized three-dimensional frontal models do not portray adequately the usual baroclinic structure of the atmosphere, and that considerable uncertainty exists regarding the connection between fronts and weather. This opinion is expressed not in the desire to do away with fronts. The forecaster who uses his fronts with restraint knows that they often serve him well. Instead, the purpose of these critical remarks is to point out the need for more research on the subject.

There are several ways in which this research may be carried out. First, more flights may be made of the type undertaken by the British. Second, if it is not too late, serial ascents might be arranged as part of the program of the International Geophysical Year. And last, but perhaps most important, by proper methods of analysis much valuable information could be gained from routine surface and upper-air observations. In particular, long series of vertical cross-sections should be prepared by some semi-objective technique such as that outlined earlier. Preconceived models should be avoided. Only a large weather organization could undertake the analysis of great numbers of cross-sections. However, now that analysis is becoming more centralized, a project of this sort should be feasible.

Before passing on to the final topic, I might mention one feature of Arctic analysis which is in some respects unique; that is, the low-level inversion which blankets the Arctic Sea in summer and the entire Arctic in winter. This inversion has been the subject of considerable research and is regarded as important in Arctic forecasting, but it is not generally represented on routine Arctic charts. A long series of carefully prepared cross-sections would appear to offer the best means of clarifying the role of the inversion in the synoptic meteorology of the Arctic.

As my last topic, I would like to discuss the question of the adequacy of the present Arctic weather network for numerical prediction. *Fig. 5* shows the number of observations available if all stations report at one time. Unfortunately, it is usual for rather large numbers of reports to be missing on any given chart. An idea of the adequacy of the Arctic network can be

gained from the results of barotropic forecasts of 500-mb height changes made by means of the FJØRTOFT method.[3] These forecasts were made only under ideal data conditions and in a region of comparatively good data. This area is shown in *Fig. 5.*

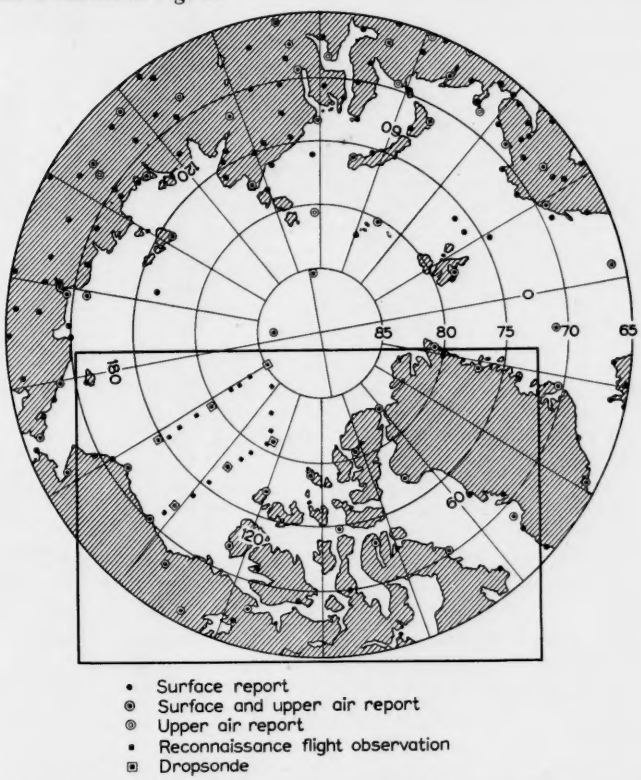


Fig. 5. Arctic weather station network in January 1956. Rectangle denotes area within which numerical forecasts were carried out

Tests were conducted over a number of five- or six-day periods, and only those periods were selected in which the "Ptarmigan" flight was entered on all maps. The results of the forecasts are shown in *Table 2.* Correlations between observed and predicted 24-hr height change range between 0·83 and 0·20. The average correlation for the 18 cases is 0·56.

Examination of individual cases suggest that the poorest correlations are due to sudden developments in the field of motion rather than to uncertainties of analysis. Although the average correlation of 0·56 is not high, the numerical forecasts had some encouraging aspects. Generally polar vortices moved in the direction of the mean flow pattern. In one instance a cold low was forecast correctly to travel first westward, then southward and finally south-eastward on successive days. To the forecaster with experience in middle-latitude forecasting such a path may seem erratic; but as was apparent from the success of the numerical prediction, it was to be explained on the elementary principle of the conservation of absolute vorticity.

Suppl.
Part.I
1956

Table 2. Summary of Correlations between 24-hour Observed and Predicted 500-mb Height Changes prepared by means of Fjørtoft Method

	Date	*Number of points*	*Correlation*
November 1955: Average correlation 0·50	1–2 November	60	0·20
	2–3 November	54	0·25
	3–4 November	55	0·64
	4–5 November	55	0·81
	5–6 November	69	0·62
December 1955: Average correlation 0·67	12–13 December	70	0·51
	13–14 December	56	0·75
	14–15 December	70	0·58
	15–16 December	54	0·82
	16–17 December	66	0·68
January 1956: Average correlation 0·50	2–3 January	66	0·47
	3–4 January	77	0·64
	4–5 January	70	0·55
	5–6 January	63	0·36
February 1956: Average correlation 0·53	5–6 February	72	0·37
	6–7 February	66	0·47
	7–8 February	69	0·83
	8–9 February	58	0·45

Average correlation for 18 cases: 0·56

Eighteen cases is a rather small number upon which to base any definite conclusions concerning the feasibility of numerical prediction in the Arctic. However, there are grounds for believing that it may be carried on with reasonable success in areas where the station density is comparable with that in the Canadian Archipelago. This includes more than 50 per cent of the area north of the Arctic Circle.

References

[1] Boville, B. W. "Meteorological Office Discussion" *Met. Mag.* Vol. 85, pp. 83–88, 1956

[2] Charney, J. G. "On a Physical Basis for Numerical Prediction of Large-scale Motions in the Atmosphere" *J. Met.* Vol. 6, pp. 371–385, 1949

[3] Fjørtoft, R. "On a Numerical Method of Integrating the Barotropic Vorticity Equation" *Tellus* Vol. 4, pp. 179–194, 1952

[4] Godson, W. L. *Informal papers in Arctic Meteorology*, Scientific Report No. 1; McGill University, Montreal, pp. 21–27, 1955

[5] Penner, C. M. "A Three-front Model for Synoptic Analyses" *Quart. J. R. met. Soc.* Vol. 81, pp. 89–92, 1955

[6] Rossby, C.-G. "Planetary Flow Patterns in the Atmosphere" *Quart. J. R. met. Soc.* Vol. 66 (Supp.), pp. 68–87, 1940

[7] Sawyer, J. S. "The Free Atmosphere in the Vicinity of Fronts" *Geophys. Mem.* No. 96, pp. 24, 1955

[8] Sawyer, J. S. and Bushby, F. H. "A Baroclinic Model Atmosphere Suitable for Numerical Integration" *J. Met.* Vol. 10, pp. 54–59, 1953

[9] Sutcliffe, R. C. "A Contribution to the Problem of Development" *Quart. J. R. met. Soc.* Vol. 73, pp. 370–383, 1947

Suppl.
Part. I
1956

THE QUANTITATIVE REPRESENTATION OF THE NORTH POLAR PRESSURE FIELD*

F. KENNETH HARE
Arctic Meteorology Research Group, McGill University, Montreal, Canada

As a prelude to research into the general circulation of the atmosphere north of latitude 55°N, the McGill Arctic Meteorology Research Group has been examining suitable techniques of quantitative description of Arctic circulation patterns. This paper examines methods of obtaining analytical expressions for the surface and 500-mb pressure fields over the north polar region. The term "specification" is used to describe this technique of quantitative representation. Suitable coordinates require careful choice, since the convergence of the meridians makes ordinary geographical coordinates useless near the Pole. Two systems of pressure sampling, corresponding to two different coordinate systems, are in operational use at McGill University. One is a Cartesian square grid, equal-area with respect to a stereographic projection of the pressure field, and the other (devised by W. L. Godson) is a spherical grid in which sampling frequency along selected colatitudes is made to vary so as to preserve the equal-area property. All systems of specification examined involve (i) the selection of suitable functions, picked either because of computational convenience or physical suitability, (ii) the sampling of the pressure field on an agreed array, and (iii) the fitting by least-squares techniques of the variable pressure to the chosen function. Labor in computation is conserved by the selection of orthogonal sets of functions, which give the normal equations of least-squares as independent equations. The earliest attempt at such specification investigated by the group was due to Haurwitz and Craig, who used a system of spherical surface harmonics unsuitable to the polar area. Two methods are in active use at McGill. The first, originally proposed by Wadsworth, Bryan and Gordon, involves the fitting of the pressure to a series of orthogonal polynomial surfaces developed by Fisher from Tschebycheff's series. This enables the McGill analysts to express 80–95 per cent of the geographical variance of pressure in the north polar area in terms of 20–30 correlation coefficients, each completely independent. The second, proposed by Godson, involves the use of Fourier transforms, in which the colatitude terms are expanded in orthogonal polynomials. Very high reductions of variance are obtained by the Godson method, using a limited number of coefficients. These two methods of quantitative representation will be used at McGill in a number of studies of the polar circulation.

1. INTRODUCTION

THE chief difficulty confronting the meteorologist who wishes to make a study of the circulation over a limited area of the earth's crust is the lack of quantitative methods of synthesis. The synoptic chart gives a useful picture

* The research reported in this paper has been supported in part by the Geophysics Research Directorate, U.S. Air Force Cambridge Research Center. The author's thanks are extended to Dr. T. F. Malone and Mr. Robert Miller, Travelers Research Center, for an introduction to this subject, and for much subsequent advice. His thanks are also due to Dr. Warren L. Godson, Canadian Meteorological Service, who devised the entirely new system described later.

of instantaneous flow, and the mean pressure chart does likewise for resultant flow, in both cases provided that the geostrophic approximation can be held valid (a provision true of polar latitudes, though not of mountainous or tropical areas). But neither the synoptic chart nor the mean-pressure map are at all easy to grasp. They show a bewildering complexity of pattern that defies effective plain-language description, and they are difficult to catalogue in any theoretically valid manner. It is natural, therefore, that the meteorologist should turn his mind to the analytical expression of the synoptic and mean-pressure patterns, in the hope that quantitative description may emerge.

As a prelude to their investigation of the polar circulation itself, the author and his colleagues at McGill University have spent much effort on finding and applying suitable techniques of *specification* to the north polar pressure field, both at sea level and at the ordinary aerological levels. The term specification is used here in the sense of WADSWORTH, BRYAN and GORDON,[11] who defined it as ". . . a mathematical representation of the distribution of some variable over a stated region of space". In our case, as in theirs, the variable is pressure; but the methods described are perfectly general, and may be applied to distributions of any other arbitrary but continuous variable on the earth's surface. A report on their performance in actual application will be published later, after machine-computation of test runs during the autumn of 1956.

It is hardly necessary to state that specification and its uses are entirely distinct from numerical weather prediction. Specification is a frankly statistical technique, and so are the uses to which it is put. The objective, however, is to demonstrate physical processes—processes that might otherwise be obscured by the almost limitless complexity of the circulation. These remarks apply with full force to the application of specification to the polar atmosphere. It is a means to an end, not an end in itself.

2. SPECIFICATION TECHNIQUES

The earliest work on specification of which the author has knowledge is that of WADSWORTH, BRYAN and GORDON,[11] to which reference has just been made. Their work, which took place between 1942 and 1946, was prompted by the search for analogues, a prediction technique that has since proved disappointing. Nevertheless, their work on specification by means of orthogonal polynomials was of lasting value, and forms the basis of one of the methods advocated below for polar areas. In 1944 HAURWITZ and CRAIG[6] perfected a method for representing analytically the pressure field over the entire earth, using spherical harmonic analysis, a method previously employed in the representation of the field of geomagnetism and the electric potential about a sphere. Since these authors reported on their work, several other groups have employed similar methods in pressure field specification, notably Malone, Miller, Friedman and their associates at the Massachusetts Institute of Technology, and the Travelers Weather Research Center. None of these groups, however, has specifically considered the polar areas, though attention is now being devoted to specification of the entire Northern Hemisphere pressure field.

A study of these and other works makes it possible to state certain steps which seem necessary in all systems of specification:

(*a*) *The selection of suitable space coordinates.*

The pressure is expressed as an arbitrary function of space coordinates in each system, but the choice of coordinates varies. WADSWORTH, BRYAN and GORDON,[11] and later BRYSON and KUHN,[1] used latitude and longitude as a curvilinear grid, ignoring the convergence of the meridians. HAURWITZ and CRAIG,[6] however, used latitude and longitude as true spherical coordinates. In the polar regions, the convergence of the meridians render the latitude-longitude grid difficult to use unless corrections are made for the convergence. For polar areas an arbitrary Cartesian grid has certain advantages that will be discussed below.

(*b*) *The selection of suitable mathematical functions of the space coordinates.*

The functions employed are chosen either because of their ease and efficiency in specification (as in the case of the orthogonal polynomials) or because they ape nature, their graphs resembling pressure patterns (as with spherical surface harmonics and Fourier series). The most successful functions fulfill both requirements. If the function is of a kind that readily lends itself to physical interpretation, the statistical process of specification ought to suggest directly the dynamics of the circulation.

(*c*) *The fitting of the variable pressure, as observed at standardized sampling points, to the chosen function.*

In general, this fitting will be carried out by the method of least-squares. Since many sampling points are required, and since the appropriate functions are usually polynomials of high degree, this is computationally a laborious business. The labor may be minimized, however, by the selection of *orthogonal* functions, a point discussed below.

To sum up, specification requires the wise choice of coordinates, suitable orthogonal functions and a least-squares fitting of the data to the functions. Even with orthogonal functions, the labor of computation involved in the fitting absolutely requires (at least for operational purposes) the use of high-speed digital computers.

3. THE CHOICE OF FUNCTIONS

Though the list of orthogonal functions is long, the limitations discussed in para. 2(*b*) effectively restrict the choice to a few. Arguing from the premise that the function should simulate observed patterns, HAURWITZ and CRAIG[6] advocated the use of spherical surface harmonics of the form

$$P_n^m(\cos\theta)\cos m\lambda \text{ and } P_n^m(\cos\theta)\sin m\lambda \qquad (1)$$

where θ = colatitude, λ = longitude, n and m are positive integers, and P_n^m is the seminormalized associated Legendre function of degree m and order n. These functions are not orthogonal over a limited latitude range, and are computationally unsuitable for the polar belt. It is possible, however, that

Suppl.
Part. I
1956

ingenious transformations may partially overcome these difficulties, in which case the Haurwitz-Craig model may well deserve re-examination.

Godson, in a paper to be published elsewhere, also uses a combination of functions that seem *a priori* physically suitable, but which are far more applicable to the polar regions. He uses Fourier expansions round selected colatitude circles, with the coefficients as functions of the colatitude expressed in terms of normalized orthogonal polynomials. Moreover, he weights the sampling grid in such a way as to correct the errors introduced by the convergence of the meridians. If some system of harmonic analysis is preferred, Godson's seems by far the most logical, and we have adopted it for general use. The computation involved is extensive, however, and the parameters required for specification are numerous. The author returns to Godson's method later.

For simple analogue determination, and for studies of development, a simpler model seems necessary. The soundest available appears to be the Wadsworth-Bryan-Gordon technique, using Fisher's orthogonal polynomials, but applied to a simple square Cartesian sampling grid. Preliminary results have shown that this method is very economical, and gives excellent results in the polar case. It must be admitted, however, that physical interpretation is much less easy than with Godson's method.

4. Specification by Means of Orthogonal Polynomials

An extended review of the use of orthogonal polynomials in pressure-pattern specification has already been given by Wadsworth, Bryan and Gordon[11] (whose treatment, slightly modified, forms the basis of the present summary) and other reviews have been circulated by Bryson and Kuhn[1] and Malone.[8] Accordingly, the present treatment is given in the shortest possible compass, allowing for the fact that none of the above reviews have been published in the meteorological journals.

Suppose that x and y are the chosen space coordinates, and that measurements of an arbitrary variable $f(x, y)$ be taken at equally spaced points along

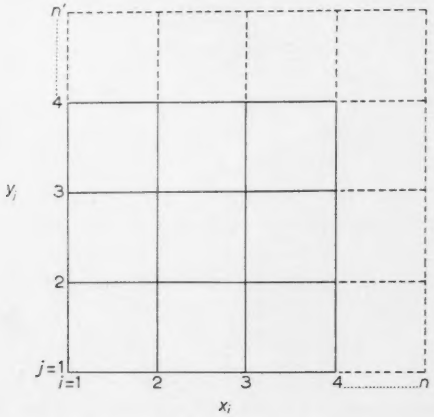


Fig. 1. The Cartesian grid notation used in the polynomial specification. The actual grid is seen in Figs. 2 and 3

$x(x_i;\ i = 1,\ 2,\ 3, \ldots n)$ and $y(y_j;\ j = 1,\ 2,\ 3,\ \ldots n')$. Then in *Fig. 1*, $f(x, y)_{ij}$ will be the value assumed by f at the intersection of column i with row j. The variable is thus specified by nn' sampled values, each $f(x, y)_{ij}$; then

$$f(x,y)_{ij} = A_0 + \sum_{r=1}^{n-1} A_r x_i^r + \sum_{s=1}^{n'-1} A_s y_j^s + \sum_{r=1}^{n-1}\sum_{s=1}^{n'-1} A_{rs} x_i^r y_j^s = \sum_{r=0}^{n-1}\sum_{s=0}^{n'-1} A_{rs} x_i^r y_j^s. \quad (2)$$

In other words, if we have nn' values of the variable, we can represent them exactly as a polynomial of degree $n-1$ in x and $n'-1$ in y, which will comprise nn' terms, each containing an unknown constant A_{rs}. If we wish to obtain an approximate representation of f, say $\hat{f}$, in terms of a polynomial of lower degree in x and y, we can do so by the method of least-squares by solving the normal equations

$$\sum_{i=1}^{n}\sum_{j=1}^{n'} x_i^r y_j^s (f - \hat{f})_{ij} = 0, \quad (3)$$

allowing r and s to go as high as is necessary to achieve the desired degree of approximation. Equations (3) are simultaneous, and their solution is hence very laborious. Moreover, they give the polynomial coefficients A_{rs}, not as the true values of the complete expansion, but as approximate values which will change every time an extra term is calculated.

In statistical work it has long been the custom in complex regression problems to use polynomial regression lines or surfaces. It is a common practice in such work to use the *orthogonal polynomials*, developed by FISHER[3] out of an original series due to TSCHEBYCHEFF.[10] The technique used by Wadsworth, Bryan and Gordon, and applied here to the north polar region, is based on the replacement of the ascending powers of x and y in Eq. (2) by functions $X(x)$ and $Y(y)$ which are orthogonal polynomials in x and y. These polynomials have been extensively tabulated by FISHER and YATES[4] and more recently by DE LURY,[2] who has taken them to the 25th power.

If, in Eq. (3), we use the raw powers of x and y, many numerically large terms will result. If, however, we use X_1, X_2, X_3 . . . etc., and the equivalent polynomials Y_1, Y_2, Y_3 . . . , equations (3) become independent equations, which may be written

$$\sum_{i=1}^{n}\sum_{j=1}^{n'} X_{ri} Y_{sj} f(x,y)_{ij} - a_{rs} \sum_{i=1}^{n}\sum_{j=1}^{n'} (X_{ri} Y_{js})^2 = 0. \quad (4)$$

The fact that Eqs. (4) are independent means that all the polynomial coefficients a_{rs} are uncorrelated, and may therefore be determined one by one without changing the values of preceding coefficients. Hence each a_{rs} has the same value as it would possess in a complete functional representation of $f(x, y)$. *In effect, if $f(x, y)$ is the field of pressure, these coefficients can be regarded as accurate parameters of the geostrophic circulation, mutually uncorrelated and exact in value.*

WADSWORTH, BRYAN and GORDON[11] modified this procedure by normalizing the coefficients a_{rs} and the polynomials themselves. The orthonormal polynomials X' are given by the relation

$$X'_{ri} = X_{ri}\left(\sum_{i=1}^{n}\sum_{j=1}^{n'} X_{ri}^2\right)^{-\frac{1}{2}}. \quad (5)$$

Furthermore, they compute a measure of the variability of $f(x, y)$ about $\bar{f}$, its own mean,

$$S = \left(\sum_{i=1}^{n} \sum_{j=1}^{n'} (f_{ij} - \bar{f})^2 \right)^{\frac{1}{2}}. \tag{6}$$

With these refinements, Eq. (2a) can be replaced by

$$\hat{f}(x, y)_{ij} = \bar{f} + \sum_{r=0}^{b} \sum_{s=0}^{c} Z_{rs} S X'_{ri} Y'_{sj}, \; b << n, \; c << n'. \tag{7}$$

In the double summation, the term in which $r = s = 0$ is omitted. The expansion of the second term on the right is taken to high enough values of r and s (say b and c) to explain any desired proportion of the variability within the sampling grid. Z_{rs} is the correlation coefficient between the topography of the $f(x, y)$ surface and the rs-orthonormal surface, and S is as defined in Eq. (6) above. The per cent reduction of sample variance is given very simply as

$$100 \left(\sum_{r=0}^{b} \sum_{s=0}^{c} Z_{rs}^2 \right).$$

To sum up, then, the method specifies the variable by means of the following parameters:

(i) The grand mean of the variable over the field, $\bar{f}$;
(ii) A measure of the variability, S, of the variable about that mean; and
(iii) A series of parameters Z_{rs} which express the correlation between the variable's topography and those of a series of normalized orthogonal polynomial surfaces.

FRIEDMAN[5] has published pictures of these polynomial surfaces, and it may help the newcomer to examine his illustrations carefully. The investigator is at liberty to choose as many row (i.e. $r = 0$) and column ($s = 0$) terms as are necessary adequately to express the components of the variability in those rows and columns, and to use enough cross-product terms ($r \neq 0$, $s \neq 0$) to express the residual or interaction variability. It is hence necessary as a preliminary step to analyze the total variability of $f(x, y)$ into x-, y- and xy-components (called column, row and interaction variability in this paper). In operational practice, of course, it is impracticable to make such a decision for each individual map; a standard distribution of the Zs between row, column and interaction terms has to be adopted for machine analysis. This point is discussed below.

5. The Problem of Coordinates and Sampling Grids in the Polar Area

The pressure fields over the north polar area present many points of difference from those of mid-latitudes, and the best choice of coordinate systems has to take account of this fact. There is also the obvious point that the meridians converge on the geographical pole (which is not, unfortunately, the normal pole of the upper-level westerlies, which are markedly asymmetrical to the geographical pole). In general, we shall choose coordinates and sampling systems or *grids* so as to achieve the following objectives:

(i) to avoid special emphasis on the geographical pole, and the convergence of polar coordinates.

(ii) To achieve, as far as possible, a sampling grid uniformly weighted as to area.

(iii) To make possible, without difficult transformations, the direct use of the most useful functions chosen under Section 3.

(iv) To preserve conformality.

It is also desirable to make the grid dense enough to achieve an adequate sampling of the pressure field. Some of the applications of these techniques have ignored points (i) and (ii). Thus the orthogonal polynomial methods described above have generally been based on a 5 or 10 degree latitude-longitude sampling grid; because the meridians converge, this means that the pressure field is sampled with increasing frequency as the pole is approached. If the circulation in high latitudes is more complex than in low, this may be a desirable consequence, but if the method is extended nearer the pole than, say, 60°N latitude, it is clear that the sampling becomes impossibly unequal.

Present procedures of the group involve two systems of coordinates, with the following properties and uses:

(i) *The Godson grid*, using modified cylindrical coordinates. The colatitude and the longitude are the working coordinates, and sampling is carried out on each colatitude circle with a frequency that almost exactly ensures equal-area sampling. This grid is, of course, used in the Godson technique of specification discussed later.

(ii) *An arbitrary polar Cartesian grid*, having two coordinates labelled x and y, with origin at the pole, the x-axis lying along the 25°W meridian, the y-axis along the 115°W. Sampling of the pressure field is carried out over a unit square grid, with sampling points spaced so that at the pole the frequency is equal to the mean frequency of observing points over the northern land areas.* This grid is thus completely equal-area with respect to the zenithal stereographic projection employed in our base map, used in *Figs. 2* and *3* of this paper. Complete conformality is assured.

In effect, Godson's grid samples on an equal-area basis with respect to the earth's sphere; the present grid does so with respect to a polar conformal projection of that sphere. The orientation of the grid is such that the diagonals run from the Siberian winter high's position to Labrador and from the Aleutian winter low to western Siberia. The sampling thus extends furthest from the pole towards the major centers of action of the mean winter pressure distribution.

6. The Program of Pressure Sampling

In order to test the methods of specification described in this report, it was necessary to select the best available surface and upper-air charts of the north polar area, and to set up sampling arrays of pressure from each chart. Published arrays of pressure did not suit our purpose. Moreover, we wished

* Actually, the frequency exceeds the actual number of pressure readings on any individual sea-level map by a wide margin, because it was based not on observations received, but on the network of synoptic stations that we supposed, in our ignorance, would be reporting. For the 500-mb maps, the grid is clearly too dense.

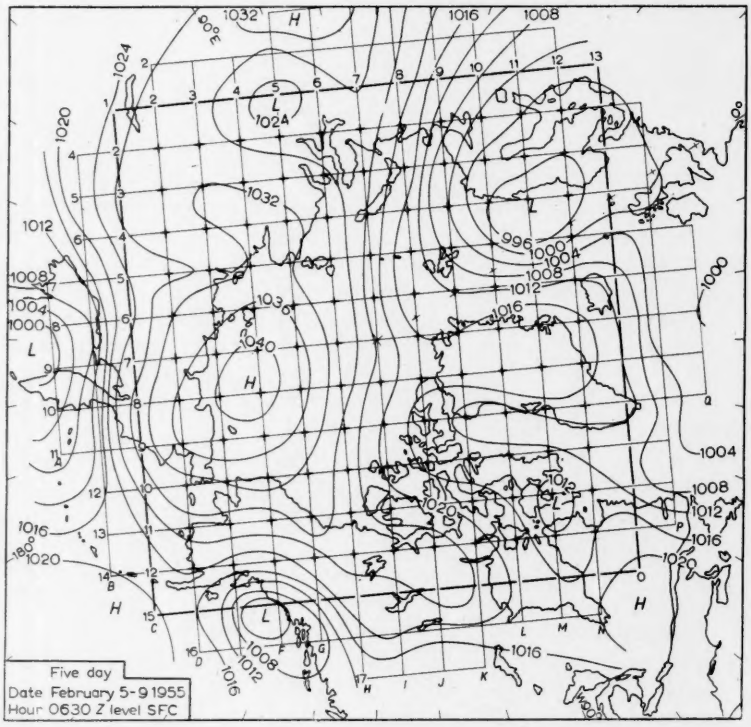


Fig. 2. Five-day mean pressure over the north polar area, 5–9 February 1955, at sea level. The 13 × 13 arbitrary grid used in specification is shown. 24 orthonormal polynomial coefficients expressed 94·62 per cent of the variance of pressure within the grid

to base our studies on the most intensive polar analysis available. Accordingly, we decided to use the chart series of the Arctic Forecast Team, Canadian Department of Transport, Meteorological Division, Edmonton, Alta. and to prepare our own sampling arrays of pressure.

The Edmonton series consists of twice-daily sea-level and 500-mb analyses of the entire Northern Hemisphere down to about latitude 50°N. The maps are prepared operationally by a group under the direction of Mr. Horace Wilson, and Ozalid copies are sent by mail to the group's office at McGill University. From them, we have prepared the following arrays, in each case for both synoptic levels:

(i) Once-daily arrays over a 13 × 13 square grid on our Cartesian system, for all 1955.

(ii) 5-day means and 30-day mean for the period January–June 1955.

These are currently being punch-carded, and will be available to all potential consumers. The square grid covers the pressure field down to 52°N along the diagonals, and to 63°N along the axes. It thus enters all three of the main winter centers of action, but stays well north of the stronger westerlies.

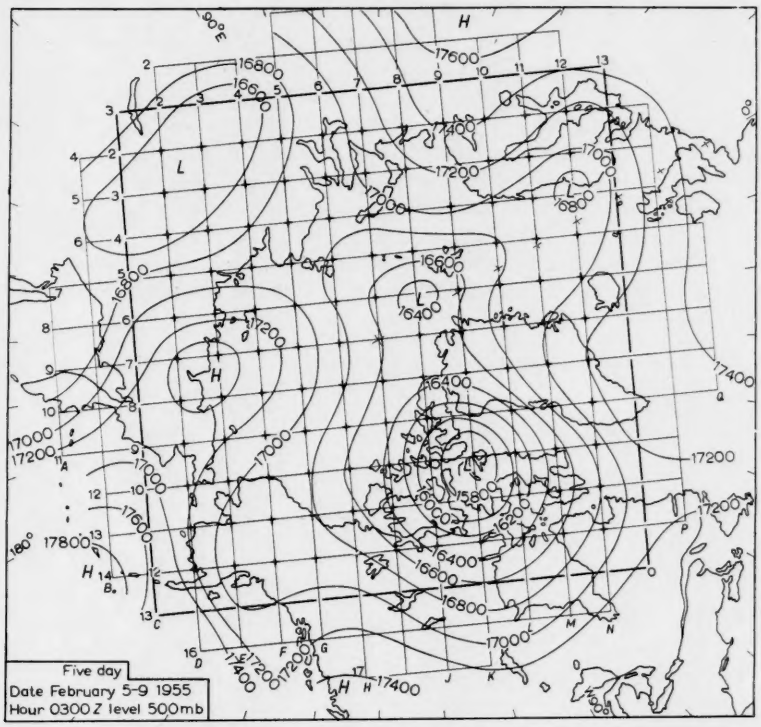


Fig. 3. Five-day mean topography of the 500-mb surface north polar area, 5–9 February 1955, corresponding to the sea-level chart in Fig. 2. 24 coefficients expressed 92·15 per cent of the variance within the grid

Arrays of pressure for the Godson grid have been prepared for a more limited number of cases.

7. Observed Variability of Pressure within Cartesian Grid

It was made clear above that the choice of row, column and interaction polynomials in the calculation of the Z-coefficients was to be based on the observed behaviour of the variability of pressure within the grid.

The pressure field in the high northern latitudes of the earth is unlike that of mid-latitudes, where the predominance of zonal flow ensures that a large part of all the variability will be meridional. Thus Friedman[5] has reported that, in the case of 700-mb circulation over central North America, Z_7, the coefficient of the linear orthonormal function along the meridians, was always high, citing cases where it accounted for over 80 per cent of the total sample variance. Similarly, Bryson and Kuhn,[1] in a study of the 500-mb topography between 10°N and 70°N, showed that the same coefficient systematically accounted for about 80 per cent of the total sample variance; even when the topography of the surface was expressed in terms of its deviation from the seasonal norm, 20–65 per cent of the variance was

Suppl.
Part.I
1956

still accounted for by the linear meridional terms. In the polar case, however, there seems likely to be no dominant "grain" to the pattern; zonal flow around the pole should be present in the outer parts of the grid, but near the pole itself experience has shown that the pattern is usually cellular. Large and small cyclonic and anticyclonic centers dominate the map, and display little of the regularity of steering typical of mid-latitudes.

Accordingly, we decided to investigate the distribution of variability as between columns, rows and interaction within our 13×13 grid. A sample of 30 cases, at roughly 10-day intervals, was selected (the length of the interval being high enough to eliminate significant auto-correlation). *Table 1* shows for both levels the mean partitioning of variability between the three possibilities.

Table 1. Mean Distribution of the Variability of Pressure, Polar Belt, between Rows, Columns and Interaction of our 13×13 Cartesian Grid

	Oct.-Mar. per cent (15 cases)	*Apr.-Sept.* per cent (15 cases)	*Annual* per cent (30 cases)	*Oct.-Mar.* per cent (15 cases)	*Apr.-Sept.* per cent (15 cases)	*Annual* per cent (30 cases)
Rows	26	23	25	20	36	28
Columns	29	24	27	33	29	31
Interaction	45	53	48	47	34	41

We must therefore expect—if this is a good sample—that in the Wadsworth-Bryan-Gordon technique the polynomials in X' and Y' will explain on the average about a half of the sea-level variance, and a little more at 500 mb. The remaining proportion of the variance—nearly half—will require the use of cross-product polynomials in $X'Y'$. We are not aware of any theory that will tell us, from the above figures, how many powers will be required in each of the three cases. Our machine calculation will, however, enable us to re-compute *Table 1* using all the cases, and it will be possible to set alongside it the performance of the individual coefficients. Thereafter it will be possible to establish a sound distribution.

8. Computation: Some Examples of the Orthonormal Coefficients

We have computed numerous examples of the orthonormal coefficients for actual synoptic cases and for 5-day and 30-day mean periods. These preliminary tests have shown that it is easy to obtain an 80 per cent reduction of sample variance, using various combinations of row, column and cross-product coefficients. It is by no means as easy, however, to attain the very high reductions reported by Friedman[5] and Bryson and Kuhn,[1] because of the absence of any clear-cut dominance by a single polynomial term.

Two examples,* chosen at random from the group's 5-day maps, are included as an illustration of the technique. The maps, reproduced as *Figs. 2* and *3*, show the normal cellular patterns of the emphatic nature typical of winter. The period is 5–9 February 1955. Case 1 refers to the sea-level chart (*Fig. 2*) and Case 2 to the 500-mb equivalent (*Fig. 3*).

* The computation of these examples was carried out by Miss Mona A. Macfarlane and Miss Cynthia Wilson. Much help was given in earlier stages by Mr. Fraser Lindsay.

Case 1

5–9 February 1955 Surface

ANALYSIS OF VARIABILITY

Component	*Sum of squares**	*Per cent*
Column	13,491·5	63·26
Row	1,716·3	8·05
Interaction	6,119·1	28·69
Total	21,326·9	100·00

Coefficients, column terms

Z_{ro} r		*Per cent reduction*
1	−0·720	51·84
2	−0·194	3·76
3	+0·266	7·08
		62·68

Coefficients, row terms

Z_{os} s		*Per cent reduction*
1	+0·102	1·04
2	−0·154	2·37
3	−0·135	1·82
4	+0·111	1·23
5	+0·035	0·12
6	−0·097	0·94
		7·52

Coefficients, interaction terms

Z_{rs} rs		*Per cent reduction*
11	+0·336	11·29
21	+0·083	0·69
12	+0·181	3·28
22	−0·040	0·16
32	−0·170	2·89
23	−0·019	0·04
31	−0·096	0·92
13	+0·009	0·01
33	−0·010	0·01
41	−0·015	0·02
14	−0·115	1·32
42	+0·149	2·22
24	+0·045	0·20
43	+0·112	1·25
34	−0·035	0·12
		24·42

Efficiency of specification: 94·62 per cent reduction of sample variance using 24 Z_{rs}

Case 2

5–9 February 1955 500 mb

ANALYSIS OF VARIABILITY

Component	*Sum of squares**	*Per cent*
Column	52,002·4	23·03
Row	26,487·0	11·73
Interaction	147,297·6	65·24
Total	225,787·0	100·00

Coefficients, column terms

Z_{rs} r		*Per cent reduction*
1	−0·034	0·12
2	+0·423	17·89
3	+0·177	3·13
4	−0·005	0·01
5	−0·108	1·17
		22·31

Coefficients, row terms

Z_{ro} s		*Per cent reduction*
1	−0·148	2·19
2	+0·098	0·96
3	+0·037	0·14
4	+0·241	5·81
5	+0·084	0·70
6	−0·103	1·06
7	−0·083	0·69
		11·55

Coefficients, interaction terms

Z_{rs} rs		*Per cent reduction*
11	−0·466	21·70
21	+0·388	15·05
12	+0·168	2·82
22	−0·124	1·54
32	−0·187	3·50
23	−0·222	4·93
31	+0·242	5·86
13	+0·056	0·31
33	−0·039	0·15
41	−0·043	0·18
14	−0·007	0·01
42	+0·150	2·25
		58·29

Efficiency of specification: 92·15 per cent reduction of sample variance using 24 Z_{rs}

* i.e. sum of squares of deviations from the grand mean over the whole sampling array.

Suppl. Part. I 1956

In each of these two cases, the analysis of variability of the raw data is given at the head. There follow the column (X'), row (Y') and interaction ($X'Y'$) coefficients, in each case with the corresponding reduction of variance. Total reductions of about 95 per cent for the surface chart and 92 per cent for the 500 mb were obtained by using 24 coefficients. As might be expected, most of the unexplained variance lay in the interaction terms, in both cases.

Arrangements have been made to run a year's trial series from the Edmonton charts, at both levels, by digital computer. Until this has been done, a final verdict on the usefulness of the method must be deferred. There is every indication, however, that something between 20 and 30 terms will suffice to bring the average reduction up to 90 per cent.

9. The Godson Technique

Reference has already been made to the technique developed by Godson as a more sophisticated alternative to the simple method discussed above. Godson's work was the direct outcome of his own critical reaction to the orthogonal polynomial method. His paper, with full tabulations of standard functions and other working constants, is being published elsewhere by the McGill Group. For the sake of completeness, however, a few words about it will be given here.

Godson set out to correct the following limitations of our first method:

(i) Its failure to sample the pressure field on a truly equal-area basis.

(ii) The difficulty of assigning physical meaning to the orthonormal coefficients.

(iii) The absence of any criterion for establishing the statistical significance of the orthonormal coefficients, or for the cessation of calculation when significance vanishes.

His remedy for limitation (i) has already been discussed: he replaces our crude Cartesian grid by a latitude-longitude grid in which the frequency of sampling along equally spaced (or nearly so) colatitude circles is made to vary so as to preserve the equal-area property on the earth's sphere, and not as in the Cartesian case, on the projection of that sphere.

He proposes, as a solution for difficulty (ii), to replace the orthonormal surfaces of the method described above by functions more likely to lead to direct physical interpretation, yet which preserve the all-important property of orthogonality. Accordingly, he derives the equation

$$\hat{f}(r_i, \theta_j) = g_i^{-\frac{1}{2}}\left[\sum_n \sum_m A_{nm} P_m(r_i) \cos n\theta_j + \sum_n \sum_m B_{nm} P_m(r_i) \sin n\theta_j\right] \qquad (8)$$

where $\hat{f}(r_i, \theta_j)$ is the estimate of the pressure at the intersection of colatitude circle r_i with longitude θ_j. $g_i^{-\frac{1}{2}}$ is a factor ensuring equal-area weighting. The two terms within the bracket are the familiar Fourier expansions along the colatitude circles, n being the wave number concerned. The coefficients A and B, however, are given, not as simple functions of colatitude, but in terms of orthogonal expansions $P_m(r_i)$ along the meridians. The expansion is carried out by using the normalized orthogonal polynomials of the method previously described, $P_m(r_i)$ being the m-th power expansion in these polynomials. The pressure field is thus specified by a series of independent

parameters A_{nm} and B_{nm}, where n is the wave number, and m the power of the polynomial employed. Direct physical visualization is possible because, in effect, the method involves the comparison of the actual circumpolar circulation with circumpolar wave-trains of varying number.

The fitting of the observed data to Eq. (8) is carried out by the method of least-squares, following essentially the same path as that already described in this paper. Godson has proposed, however, to introduce an objective criterion for the cessation of calculation—limitation (iii) above—when the "return" from the coefficients falls below a given level of significance. This point is too complex to merit discussion in this paper, and seems, incidentally, to have received little attention in the statistical literature.

We have applied Godson's method to the polar area with highly successful results, though the labour involved is heavy. Very high reductions of sample variance—of the same order as those reported for the Wadsworth-Bryan-Gordon technique—have been obtained for both surface and 500-mb maps. We intend to apply machine methods to provide a quantitative estimate of the method's efficiency.

10. Application

Specification of the polar pressure field has not been investigated merely as a mathematical exercise. The pioneer work of Malone and his colleagues has shown that specification can be employed in new and valuable climatological techniques, as well as in the perennial problems of the synoptician. We shall end this paper with a brief résumé of the uses to which the specification methods are being (or will be) put at the McGill center.

First comes synoptic climatology. During and after the Second World War it became customary to use crude indices of the circulation as independent variables against which the mass of surface weather data might be sorted. Thus summaries of cloud-base, visibility and humidities were often prepared as functions of surface wind direction, or, more elegantly, of the gradient wind direction. The term *synoptic climatology* became current for such work. More recently, Miller and Malone[7, 8] have greatly refined this concept by investigating the behaviour of surface weather elements as functions of the circulation over a large surrounding area, using the Z-coefficients as parameters of the circulation. Friedman[5] has even shown that precipitation *amounts* can be objectively predicted in this way. The McGill group therefore plans to use similar techniques to investigate north polar temperature anomalies, and other climatic phenomena under obvious dynamical control. As a special case of this method, they have already begun to classify circulation patterns over the north polar area by the analogue method, for which specification was originally designed.

A second, more fundamental investigation, will take development of the polar pressure field as its problem. Miller and Malone[7] have shown that in mid-latitudes one can predict the interdiurnal pressure change by using linear operators derived by regression techniques from the Z-coefficients. Moreover, the interdiurnal correlation between synoptic charts is readily expressible once the pressure field is adequately specified. We hope to use this property in studies of both persistence and change of type.

Lastly, we believe that these methods—and especially Godson's—can be

useful in the study of the general circulation over the polar cap. Much work remains to be done before we can justify this belief. This work is now under way.

References

[1] Bryson, Reid A. and Kuhn, Peter M. "Half-hemispheric 500-mb Topography Description by means of Orthogonal Polynomials, Part I, Computation" *Scientific Report No. 4, Air Force Cambridge Research Center, TN-56-260, U.S.A.F. Contract AF 19 (604)-992.* Department of Met., University of Wisconsin, pp. 15 with appendices unpaged, 1956

[2] De Lury, D. B. *Values and Integrals of the Orthogonal Polynomials up to n = 26;* Toronto University Press, pp. 33, 1950

[3] Fisher, Sir Ronald A. *Statistical Methods for Research Workers,* 12th. ed., pp. 147–156; Oliver and Boyd, London, 1954

[4] Fisher, Sir Ronald A. and Yates, Frank. *Statistical Tables for Biological, Agricultural and Medical Research,* 4th ed., pp. 27–29 and Table XXIII; Oliver and Boyd, London, 1953

[5] Friedman, Don G. "Specification of Temperature and Precipitation in Terms of Circulation Patterns" *J. Met.* Vol. 12, pp. 428–435, 1955

[6] Haurwitz, B. and Craig, R. A. "Atmospheric Flow Patterns and their Representation by Spherical-surface Harmonics" *Geophysical Research Paper 14;* Air Force Cambridge Research Center, pp. 79, 1952

[7] Miller, R. G. and Malone, T. F. "The Application of Synoptic Climatology to the Prediction of Sea-level Circulation Patterns" *Studies in Synoptic Climatology,* final report, Synoptic Climatology Project No. N5ori–07883, pp. 115–122, 1956

[8] Malone, T. F. "Synoptic Climatology as an Aid in Weather Prediction" To be published

[9] Russell, Pamela and Ballabon, M. B. (Editors) "Informal papers ln Arctic Meteorology" *Scientific Report No. 1, Air Force Cambridge Research Center, TN-55-896, U.S.A.F. Contract AF 19(604)-1141.* Arctic Meteorology Research Group, McGill University, pp. 36–40, 1955

[10] Tschebycheff, P. L. *Theorie der congruenzen Elemente der Zahlentheorie,* Berlin, trs. ed. by H. Schopera. pp. 31, 1889. *Not examined by author*

[11] Wadsworth, G. P., Bryan, J. G. and Gordon, C. H. "Short-range and Extended Forecasting by Statistical Methods" *Air Weather Service Technical Report* No. 05-38, 207, 1948

Addendum

This research has now been completed, and both the Godson and Wadsworth parameters applied in specific studies of the circulation now being published elsewhere. $36Z_{rs}$ at sea-level, and 25 at 500-mb, account for 90% of the variance; about 70 Godson coefficients, associated with a maximum of $n=7$, do the same.

DISCUSSION, 4 JULY 1956

Chairman: R. C. SUTCLIFFE, United Kingdom

NAMIAS: A number of interesting points have been raised today. Dr. Scherhag reported on a number of interesting items. One was the motion of the cold air droplets, or cold lows, which have always been a source of difficulty in weather forecasting. He related the motion to the velocity field surrounding the cold low aloft, in the vicinity of the 500-mb surface. That reminded me of some work that I have done much earlier with warm anticyclones. Some of you may remember from the chapter which I had in Petterssen's book, a suggestion that the anticyclonic vorticities, south of the westerlies in summer at any rate, seemed to have a tendency to move in the direction of the weakest winds, surrounding the warm anticyclone at the isentropic surface. And that is Dr. Scherhag's result that the cold cyclones tended to move in the direction of the strongest winds. In the case of the anticyclone, I have been trying to think of an explanation without success. At that time, in the 1930s, there was a good deal of talk about the non-gradient components and the frictional drive of the westerlies which would then result in a convergence of air into certain areas leading to the anticyclones. And I felt that the convergence would take place at least in the area of the least winds. So that the mass of air would drift with its anticyclones towards the area of least winds. I was wondering if Dr. Scherhag had speculated on the physical explanation of some of the results that he has shown.

SCHERHAG: I have not considered the movement of high-pressure systems, but only the movement of certain cold lows. But I believe that all basic systems will move in the direction of the strongest wind speed. The rule that the cold air droplets which do not extend down to the surface, will move along the surface isobars, is nearly the same rule, because the surface wind is frequently a prevailing wind in the upper air too. But with respect to the movement of the high-pressure systems, could it not be that we have to expect pressure fall in the area against which the strongest wind is blowing, so that the high-pressure system cannot move in that direction. But I have never reflected on this point.

FRITZ: I want to ask about the Berlin warming. At the time that the Berlin warming was discovered, several people associated this with a solar event which occurred near that period of time. I would like to ask Dr. Scherhag if he thinks there is any connection between the solar event and the Berlin warming. If not, does he have any evidence for a solar event connected with the Canadian warming, which in turn later leads to the Berlin warming?

SCHERHAG: There was an event like a solar flare, I believe, the 5 or 6 February. There was, as I know, a very strong increase in the cosmic rays about 4 or 5 February, so that it is possible that this warming in Canada was caused by solar phenomena. But, on the other hand, it is possible to explain the Berlin warming by advection, though it seems to me that it is very difficult to explain it by advection alone. We got such high temperatures over Berlin as were never observed earlier. In order to find out whether such a solar flare will have any influence on the tropospheric conditions, we have selected 22 of the strongest solar flares of the last five years. The result, not yet finished, seems to indicate, when we consider the whole area between 60°N and the Pole, a pressure fall 2 days before such a solar flare and then a steady pressure rise until 5 days later. The amount of this pressure rise

over the whole area is about 1 mb. And then follows a pressure fall again. If we had only this pressure rise, I would consider it as a real phenomena, but there is another pressure rise 7 days later, so that on the tenth day the pressure is still the same as on the fifth day after these phenomena. And therefore I am in doubt whether it is a real phenomenon or only an accident.

SUTCLIFFE: I would like to be quite clear whether it is imagined that the effect of a solar flare is through ordinary radiation or through corpuscular radiation from the sun. In the case of ordinary radiation, it is extremely hard to imagine how it would have a localized effect on some part of the atmosphere to produce a small heated area. But of course it is possible if there were some control by the earth's magnetism.

SCHERHAG: With respect to this Berlin warming, it is a fact that the warming began before the corpuscular radiation had reached the earth. If there is any connection, it has to be through the ultra-violet radiation, and not through corpuscular radiation.

It is a difficult problem to explain. It seems that the ozone-content is much lower in the extremely cold portions of the atmosphere than in other portions. It is surprising that we always have the same phenomena at the end of the winter and in early spring. This confirms earlier investigations in which connections between solar events and the pressure distribution at the surface are found in winter-time only. I believe we must devote much research work in studying the situation and the ozone-content in these cold air poles, in order to find out whether it is possible that in this cold region the ultra-violet radiation can reach much deeper into the atmosphere than in other regions.

FJØRTOFT: I am curious to know whether it is only rises in temperature and in pressure which Dr. Scherhag has observed, or do you also in the high stratosphere observe sudden cooling and sudden drop in pressure?

SCHERHAG: Cooling is observed but the cooling occurs only gradually. This Berlin warming was experienced within two days. From one ascent to the next there was a warming of about 30° or 40°C at 10 mb. And then followed a gradual cooling and decrease in pressure which continued for 14 days. I have never experienced a case with sudden cooling comparable with this sudden warming.

FJØRTOFT: If the sudden warming is due to advection, would you not think that the mechanism would work both ways?

It is very difficult, I would say, to determine whether the warming can be due to advection or not, because the dynamical mechanism is so complicated. We know, for instance, that it is possible to get a change in pressure over a large area because of a tendency for systems of smaller scales to gather in certain regions. And this would certainly result in vertical motions also. There are several dynamical mechanisms which can be imagined as possible explanations of a sudden change in temperature. I think one should try to calculate these by using the equations, and leaving variations in radiation out of the question to see if you then can get this warming. Are the data good enough at those levels so that a reliable analysis of the situation can be carried out to make physical predictions possible?

SUTCLIFFE: Dr. Fjørtoft asks a specific question: do we think there are enough data available for these higher levels to enable dynamical calculation to be carried out?

SCHERHAG: No, I don't think so. The warming at the Arctic stations was at the 50-mb level. The strongest warming over Berlin was at 10 mb, a level from which the data are very sparse.

FJØRTOFT: We cannot exclude the possibility that these rises in the very high levels are coupled with something that goes on at lower levels where there are more and better data. So one could try to make a calculation by solving the physical equations, knowing only the data for lower levels and predict what the changes would be at higher levels.

SCHERHAG: At first I had thought that we could explain all these changes in the stratosphere by changes in the troposphere. This was, however, the first case I experienced where strong changes in the upper stratosphere occurred with no changes in the troposphere. So far it is completely different from other cases.

REED: Dr. Fletcher has mentioned that high surface winds are one of the major forecast problems in the Arctic. At the University of Washington we have begun a study on this problem, the preliminary results of which I would like to present.

From maps prepared twice daily in the Department of Meteorology, a record was made of each instance in which a weather station located in northern Alaska, northern Canada or Greenland reported a wind of force 6 (28–32 knots) or greater. A tabulation of 8 months of data for the months August 1955–March 1956 revealed a surprising number of cases of high winds as defined in the foregoing manner; for example, 75 cases of force 6 winds, 29 cases of force 7, 15 cases of force 8, etc.

The first result of interest is that nearly all stations in the areas under consideration recorded high winds on one or more occasions. Moreover, Thule, Greenland, which is noted for its gales, did not have the greatest frequency during the period under investigation. Alert, Resolute and Isachsen had nearly twice as many occasions on which the wind speed was force 6 or greater.

Next, a comparison was made between actual and geostrophic winds for the cases of high speed. The geostrophic values were measured from analyses prepared by the Extended Forecast Section of the U.S. Weather Bureau. In making such a comparison there is the danger that the analyst may have spaced the isobars with the geostrophic principle in mind. However, from an inspection of the maps it appeared that the pressure gradient at any spot was determined more from pressures at surrounding stations than from the reported wind.

The comparison showed that for winds of force 6, the average geostrophic speed was 27 knots; for force 7, 29 knots; for force 8, 36 knots; and for force 9, 47 knots. All cases of speeds in excess of force 9 occurred at Alert, Ellesmere Island, and here the reported speeds were considerably greater than the geostrophic.

The tentative conclusion is that high winds in the Arctic are not due entirely to local effects. The large-scale pressure gradient is in most instances essential to their formation, and whether or not a strong gradient leads to strong winds at a particular spot depends on the interplay of the gradient wind and the local topography.

FLETCHER: I know that the research being conducted at Washington is directed towards this particular sort of thing, i.e. relating detailed occurrences with broad-scale features. I feel that the sparsity of data has quite a bearing on the matter of our inability to predict some of these small-scale features satisfactorily. I don't know how the geostrophic or gradient winds at the different stations were determined by your research staff. Analysts, though, usually try to space the isobars in such a way that they agree with the winds, especially in areas where data are sparse.

The case of the airport at Thule, which I mentioned in my paper, is one of our most important examples. This is a problem which has been plaguing our forecasters through the years since the station was established. As I pointed out, the winds mainly occur not at the base but rather away from the base in the critical approach area. The first attack in forecasting this particular phenomenon, and which was suggested to the Thule forecasters for investigation, included measuring the pressure gradients from the data of all stations which were available in the area. The attack incorporated not only measuring the gradients objectively and then attempting to correlate the gradients, station-pair by station-pair, with the strong wind occurrences; but also taking the maps (which had a considerable amount of subjectivity built into them), measuring the pressure gradients, and comparing these with the winds. There was no success with this attack.

HESSELBERG: In such a country as Norway we have in our valleys and everywhere

Suppl.
Part. I
1956

in the mountains winds that do not agree with the pressure distribution. That is caused by the effect of mountains nearby or far away, effects of the coast, and so on. My point is that many things happen between our stations and that there might be real local pressure distributions present explaining the local winds.

HARE: I think that gravity winds are overdone in meteorological literature, and that many of the local winds that are ascribed to steep slope are in fact unconnected with that slope. I remember testing one particular mistral on the north of the Mediterranean, which was ascribed to a steep slope under radiation conditions. Because there happened to be synoptic stations at both ends of the gap in the mountains, through which it poured, and an anemometer at the mouth of it, I correlated the wind speed at the station (which was Genoa) with the pressure difference across the mountains, and came up with a highly significant coefficient, something over 0·7. The inner stations were quite a long way from the mouth, and it seemed to me to indicate that the pressure difference was the thing to take account of. On the other hand, of course, the pressure difference might have been correlated with clear skies. So I did it separately with clear skies and overcast skies and got no difference at all. I do not think that that particular mistral has anything to do with radiative cooling of the slopes. But in these Arctic stations the questions of slopes might still be quite important, even if we discount gravity.

I would like to ask if there are known cases of intense local winds in the perfectly flat Arctic. Because if you have mountains, it is not only the question of the slope alone, you also have this special local turbulent effect. So that you cannot really be sure what you are dealing with; you cannot be sure what the slope is doing. Possibly it is producing slope winds, but possibly also it is setting up conditions favourable for the local turbulent effect.

KEEGAN: I think I can give a partial answer to Dr. Hare's question. It refers back to the comparison of the two analyses that I mentioned the other day. It seems from the few months investigated that the wind direction at T-3 is very significant. I did not measure just how geostrophic or gradient it was in velocity but, as far as direction goes, I am of the opinion that any wind reported in the central Arctic Basin should be accepted for the drawing of isobars.

FLOHN: May I raise another point, which has not only importance for the Arctic area but which is rather general, and depends upon what Dr. Reed has said. In the present stage of development of our science it is generally felt that some classical textbook concepts should now be revised according to our recent experiences in describing the facts as well as regarding the differential equations used in numerical prediction. And from this viewpoint it seemed to be necessary to revise our concepts of synoptic analysis too, as pointed out by Dr. Reed. What does the forecaster really need? I think the real need is not a representation of air-masses and fronts, but an exact representation of the temperature field, of the wind field, of the pressure field, so far as the geostrophic approximation is sufficient for routine practice, and of the fields of the cloudiness and precipitation. And it is not clear that we can fill these needs *completely* by the usual frontal and air-mass concept. I feel it will be necessary to discuss this point very thoroughly in the near future. I am not sure if a radical change in analysis methods would be advisable. But we should put radical questions as a basis for discussion, when we may reach a useful compromise between a modern type of upper-air analysis and the classical methods.

NAMIAS: I think it is quite apparent that there are certain classes of systems which may not appear to fit in with the frontal model. But this circumstance would be bound to occur with almost any model that was established on incomplete data. For example, smaller scale systems, such as the squall-line or pressure-jump-line, frequently move ahead of the cold front and at a different speed. This new concept does not, however, invalidate the general cyclone model with its cold front, warm

front and occlusion. It is rather an addition. Another similar modification arose as early as 1926 when J. Bjerknes pointed out the fact that the backbent occlusion is not in the ordinary sense an occlusion, but a reflection of the upper level which had properties like a front. As time proceeds, we are getting a clearer picture of the cloud and hydrometeor forms and their relation to various atmospheric circulation systems. These new discoveries in no way detract from the importance of the classical model and the associated family of cyclones which is now known to be related to the long planetary waves. There is in meteorology a tendency for each generation to forget or criticize the work of the more immediate earlier generations, thinking that the new attack is revolutionary and completely incompatible with older ideas. I think we would make more progress by building upon and utilizing the work of earlier generations rather than by trying to displace it entirely. While no one believes the concept of fronts and air-masses provides the complete answers to weather problems, it cannot be denied they remain a vital part of weather analysis. I find it difficult to conceive the time when a forecaster will ever abandon completely the ideas of frontal structure.

SUTCLIFFE: I feel it is an important subject. The statements made by Dr. Reed rather indicated that the amount of agreement between the analyses by different services was extremely poor. My impression is, provided that the fronts are well-marked, good fronts, that we get a very good measure of agreement. Fronts are not theories, fronts are states of the atmosphere, just as real as thunderstorms. We cannot do without them. They were discovered and they exist, so there is no question of getting rid of them. The difficulties come in because fronts are created and fronts disappear, so that there must be many weak fronts, which are either growing or disappearing or never really come to anything, which are below the threshold of one analyst's criteria, although another analyst may accept them. And without objective criteria I do not think we can ever remove this discrepancy with weak fronts. But still I think there is a lot of room for clarification. We should have some other ways of indicating rain areas which develop perhaps by vertical instability or in other ways which really are not a matter of air-mass boundaries.

HESSELBERG: I was very glad to hear the two last speakers defending fronts. All old meteorologists, who have worked with them, know they are realities.

REED: Dr. Namias spoke of the necessity for each generation to build on the work of the previous. And I think that the history of science shows that this is necessary for progress. Moreover, I think that criticism just for criticism's sake is not desirable. However, if we look more closely at the growth of scientific knowledge I feel that we shall find that the building is not an orderly process but that it proceeds by fits and starts and that there is often much to be gained by examining an old problem or concept in a new light or from a somewhat different point of view.

So I hope that my remarks of this morning will not be interpreted as an attempt to tear down the frontal concept but rather as a plea for a fresh look at the problem. In my paper I said that I believed in the existence of fronts. So there is no disagreement in this respect. Also, I agree with Dr. Sutcliffe that sharp fronts will be analysed much the same by different analysts.

My reservations regarding the classical frontal model are mainly derived from experiences gained in cross-section analysis. The Norwegian model appears especially valid in the analysis of weather systems over the Pacific Ocean during periods of high index. However, the difficulty comes when we attempt to verify the model in three dimensions by means of aerological data.

FLOHN: In most cases the Norwegian concepts of fronts will be confirmed, but there are other situations in other regions, especially in subtropical latitudes and in connection with large-scale blocking, where this is not the case. There are cold drops, upper troughs, etc. which are not well represented by surface fronts. In those cases, I feel it would be advisable to improve analysis technique.

Suppl.
Part. I
1956

AN EMPIRICAL METHOD OF FORECASTING 24-HOUR MOVEMENT AND CHANGE OF INTENSITY OF SUB-ARCTIC NORTH ATLANTIC CYCLONES*

J. J. George, P. M. Wolff and W. L. Somervell, Jr.
Bureau of Aeronautics Project AROWA, U.S. Navy

A method is presented for objectively predicting the 24-hr movement and change in intensity of North Atlantic cyclones. The technique developed requires only measurement of the 500-mb height and temperature gradients above the current sea-level center, and determination of the type of 500-mb flow in which the sea-level system is embedded.

Although the results, which are empirical, speak for themselves, the success of the method is primarily attributed to proper classification of cyclones and to weak low-level baroclinity over the oceans, where the sea surface tends to create relatively uniform temperature gradients in the lower levels. That the 500-mb analysis alone should be sufficient to define the movement and intensification characteristics of maritime cyclones is also ascribed to the following: (1) since the 500-mb height gradient is measured at a point above the sea-level center, the strength of the 500-mb flow is equivalent to the sea-level 500-mb thermal wind; (2) inasmuch as 500-mb temperatures are closely related to 500–300 mb mean temperatures, 500-mb isotherms are an excellent approximation of the 500–300 mb thermal wind; and (3) since the geostrophic wind direction between 500 mb and the tropopause seldom varies appreciably from that at 500 mb when the maximum wind in the layer is greater than 50 knots, the 500-mb contours adequately represent the direction of flow in the middle and upper troposphere.

In both the intensification and movement forecasts sea-level lows are classified as to type of 500-mb flow in which embedded and location therein, effectively classifying individual lows as to stage of development. It has been verified that movement and change of intensity are closely related, for the most strongly deepening storms usually move rapidly and filling storms as a rule move slowly.

Due to the strong low-level baroclinity over continents, large errors are more often encountered with storms originating over North America than in those over the North Atlantic, or even the more maritime western Europe. Nevertheless, the method is useful over all three areas.

The change of intensity forecast gives best results during the winter season, and is particularly useful in anticipating the explosive deepening which results in major ocean cyclones. Predictions of speed and direction of movement have also been tested only during the winter months, October through March.

* The full paper is likely to be published elsewhere, but no reference is yet available.

SECTION IV

Suppl.
Part.I
1956

Supp
Part
1956

SOLAR RADIATION MEASUREMENTS IN THE ARCTIC OCEAN

SIGMUND FRITZ
U.S. Weather Bureau, Washington, D.C.

Measurements of solar radiation from the sun and sky on a horizontal surface have been made on the ice island, T-3, by the U.S. Air Force Cambridge Research Center. The results for the year 1953 are summarized and compared with similar measurements in the Antarctic and in middle latitudes.

For overcast days in mid-summer the amount of energy received is similar to that received in the Antarctic. When the data are compared to those for Washington, D.C., the fraction of the extra-terrestrial energy which is transmitted to the ground is greater in the Arctic. This may be explained by the higher surface albedo in the Arctic, but differences in cloud properties may also contribute to the explanation.

The relation between the measured solar energy and the amount of ice melted is briefly discussed.

uppl.
art.I
956

INTRODUCTION

FOR several years the U.S. Air Force Cambridge Research Center has made various measurements on ice islands in the Arctic Ocean. The amount of solar energy received on a horizontal surface from sun and sky was among the measurements, and this report summarizes the data taken on the ice island T-3 during the year 1953.

During the period from March through October when the solar energy was measurable, T-3 was located near 86°N and 90°W, which is in the Arctic Ocean north of Hudson Bay, Canada. During this period it actually crossed the 90th meridian, moving slowly toward the east.

INSTRUMENTAL ERRORS

The measurements were made with Eppley horizontal-surface pyrheliometers. In polar regions, horizontal surface instruments have two main sources of error.[1] One is the so-called angle of incidence error; when the sun is low above the horizon, and the sky is clear, these instruments do not record a high enough radiation. The effect is possibly caused by reflection from the instrument receiving surface, so that when the sun is low not all the solar energy is absorbed; when the sun is high the instruments record correct values. According to laboratory tests at the Weather Bureau on a few types of instruments, all the instruments tested displayed an angle of incidence effect, although not all to the same degree, or even in the same direction.

It should be noted, however, that when the sky is overcast very little error is to be expected from this source because most of the energy comes from well above the horizon where the angle of incidence error is quite small.

Additional error may be introduced by the "temperature effect". Laboratory tests show that the pyrheliometers record somewhat more energy when the temperature is low than when the temperature is high. This amounts to about 1–2 per cent for a 10°C change in ambient temperature. Since most pyrheliometers are calibrated at temperatures near 20–30°C, an error of about 2–6 per cent may be expected when the air temperature is near 0°C. It is relatively simple to correct for the temperature effect if the pyrheliometer has been calibrated and the air temperature is available. Although no corrections have yet been made to the T-3 data, the temperature effects will be small by comparison with the major effects such as cloudiness and surface albedo.

Data

In mid-summer, at least, the errors are small because the sky is mostly overcast; the amount of cloudiness is shown in *Fig. 1.* This figure shows that during July, for example, the sky was always covered with at least 7/10 clouds and on 27 days out of 31 the cloud cover was 9/10 or more. Similar conditions also prevailed during May, June, August and September; during March and April several days with little cloudiness appeared. With so many overcast days from May through September, we should expect little error due to the angle of incidence effect.

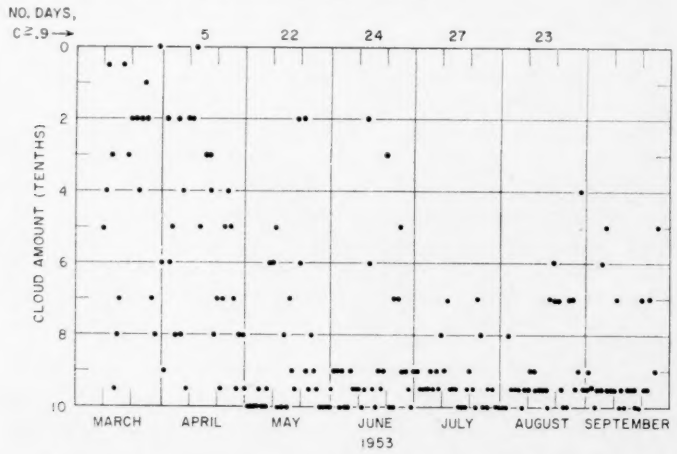


Fig. 1. Daily average amount of clouds (average of 3-hourly observations)

With that introduction, the basic solar data are given in *Fig. 2.* Here we see that the radiation falls to less than 25 ly/day when the sun is below the horizon (except for refraction) and increases rapidly as the sun rises and the length of day increases (one ly = one langley = 1 cal/cm²). The highest value recorded was about 890 ly/day in mid-June, and compares with the highest value of about 680 in June 1953 at Washington, D.C., near latitude 39°N. The larger value in the Arctic is due mainly to the 24-hr daylight in that region, and to the higher surface reflectivity.

If we compare the energy received with the extra-terrestrial (ETR)

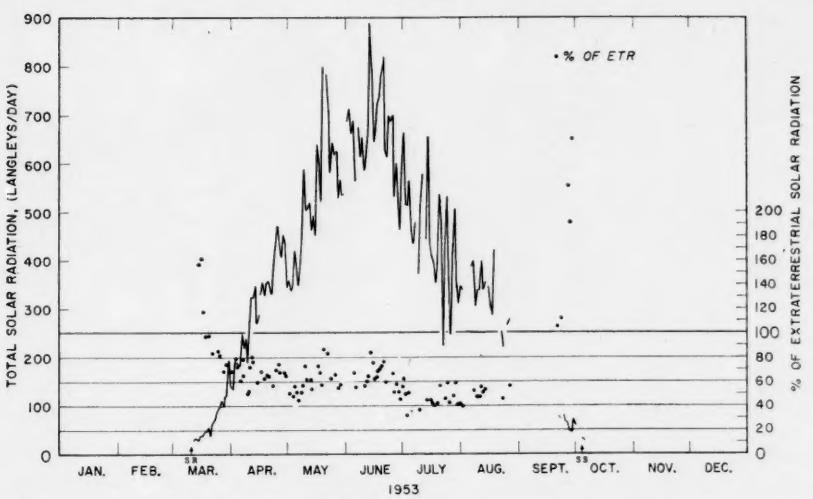


Fig. 2. Daily totals of observed radiation from sun and sky on a horizontal surface. Dots represent percentage of extra-terrestrial solar radiation on a horizontal surface

amount received on a horizontal surface outside the earth's atmosphere, *Fig. 2* also shows that in April, May and June the energy received is about 65 per cent of the ETR, while during July and August it is more nearly 50 per cent. When the energy gets very low in March, September and October, the fractions of ETR received are often well over 100 per cent. This is doubtless due to refraction of the solar beam, and to scattering of light by the air and clouds when the sun is near the horizon. When the sun is below the horizon, the expected ETR is zero, and if any energy is measured at all, the fraction of ETR actually becomes infinite. The variations in mid-summer, where this abnormality does not exist, will be discussed later.

It is interesting to compare the T-3 data with Antarctic data. LILJEQUIST[2] has published some measurements for Maudheim near 71°S latitude. A comparison between the two stations would obviously depend on the relative cloudiness. Fortunately, Liljequist segregated his data for overcast days and published the results separately. Since over the Arctic during summer the sky is overcast most of the time, data for overcast days were plentiful and these could be compared with Liljequist. Such a comparison is shown in *Fig. 3.*

Here we see first a comparison of the ETR. In mid-summer, both stations have about the same ETR but, in early spring and in autumn, Maudheim has significantly higher ETR because it is farther from the pole than T-3 was. The data measured in both places for overcast sky conditions are also shown in *Fig. 3.* Here we see that Maudheim tended to be higher than T-3 except in mid-June. We can also examine those days when the cloudiness at T-3 was 9+, i.e. between 9 and 10 tenths. Even though the T-3 radiation data is now somewhat greater than for 10/10, the comparison with Maudheim is still about the same.

As the amount of cloudiness decreases, we should expect the radiation to

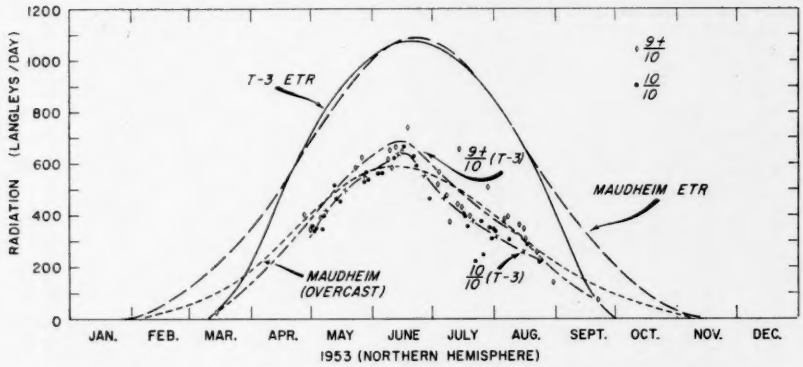
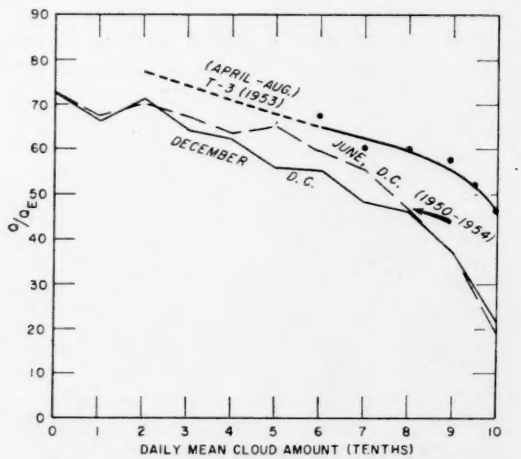


Fig. 3. Comparison of solar radiation on T-3 with that on Maudheim for overcast days

increase; the relation between cloud amount and radiation is shown in *Fig. 4*. Unfortunately, in mid-summer, when the data are most reliable, the sky is mostly overcast in the Arctic (*Fig. 1*); no cloudless days occurred at all in 1953 from May through September. Thus the relationship between cloud amount and radiation is most reliable for cloudiness of 8/10 or more. *Fig. 4* shows this relationship. In *Fig. 4*, Q is the radiation received on any one day, and Q_E is the ETR calculated for the same day; the ratio Q/Q_E normalizes the radiation to a common base. For comparison the same information is shown for Washington, D.C., for the years 1950–54. For overcast days, we note that the transmission through T-3 clouds is more than twice that through Washington clouds. The solar elevation can probably not affect the results appreciably, because in December in Washington the solar elevation at mid-day is similar to that at T-3 in mid-summer.



Variation of solar radiation on T-3 and in Washington, D.C., as a function of cloud amount. Daily radiation expressed as fraction of extra-terrestrial radiation

Clouds and Surface Albedo

The radiation data can serve as a means for (1) analyzing cloud characteristics and (2) discussing energy in relation to ice. The difference in transmission between Washington and T-3 on overcast days (*Fig. 4*) is doubtless to be sought in the cloud characteristics and/or in the albedo of the underlying surface. If we take the energy Q_0 falling on the cloud top to be about 80 per cent of Q_E and neglect absorption, we can get a first approximation to the reasons for the difference. Since $Q/Q_E = 0{\cdot}45$ at T-3 according to *Fig. 4*, $Q/Q_0 = 0{\cdot}54$. On the basis of theoretical computations (Fritz[3]) a relation between cloud parameters and surface albedo for $\zeta = 60°$ is shown in *Fig. 5*. This graph shows the relation between the cloud's optical thickness, h/L, and the surface albedo when the transmittance through a cloud is 0·54. In the vertical ordinate h is the geometric thickness of the cloud and L is the mean-free-path of the light through the cloud; L depends on the drop-size and liquid-water content of the cloud. Thus, if we consider a cloud with a given amount of liquid water per cm³ and a given drop-size distribution, h/L can be considered as the geometric height in relative units. The surface albedo was not measured in 1953. But in August 1955 it was 0·75 over the ice on T-3. From *Fig. 5*, we see that for $\tau = Q/Q_0 = 0{\cdot}54$, if the surface albedo were 0·75 the cloud "thickness", h/L would be 11 units. At Washington, according to *Fig. 4*, $Q/Q_E = 0{\cdot}2$ or $Q/Q_0 = 0{\cdot}2 \times 1{\cdot}2 = 0{\cdot}24$. A curve for this value of $\tau = 0{\cdot}24$ is also given in *Fig. 5*; from this curve we

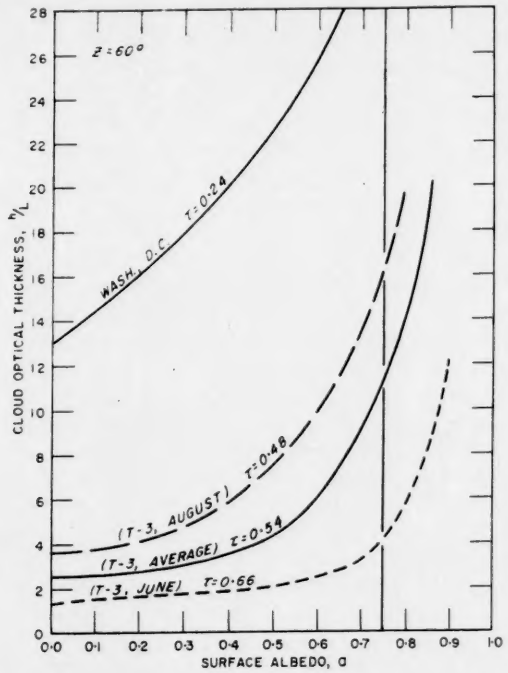


Fig. 5. Relation between cloud optical thickness and surface albedo for given cloud transmissivities, τ, when sun's zenith distance is 60°

Suppl.
Part. I
1956

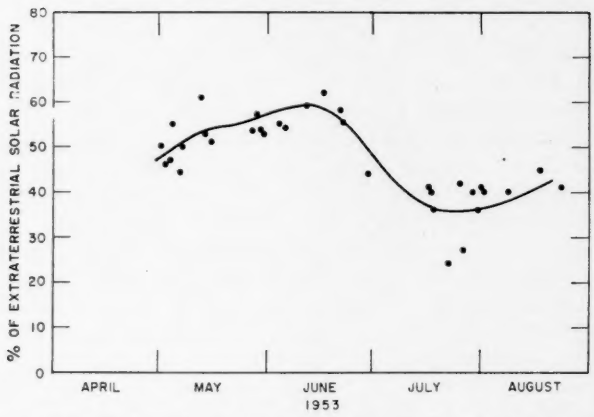


Fig. 6. Time variation of solar radiation on overcast days on T-3, expressed as a fraction of extra-terrestrial radiation

see that if we take a for Washington about 0·2 to account for some of the winter snows, h/L would be 16 units. Thus, the optical cloud thickness is about 1·5 times greater at Washington than at T-3. Of course, if the surface albedo at T-3 was different from 0·75 during 1953, this result would have to be changed accordingly. For example, if $a = 0{\cdot}7$ at T-3, the clouds at T-3 would have about half the optical thickness, although not necessarily half the geometric thickness, of the Washington overcast clouds; if the surface albedo at T-3 was 0·82, then clouds at Washington and T-3 would have had the same optical thickness.

This same line of reasoning can be applied to examine the variation in cloud characteristics or surface albedo throughout the summer. *Fig. 6* shows Q/Q_E for overcast days. There seems to have been a downward trend in Q/Q_E especially from June to September. If we take $Q/Q_E = 0{\cdot}55$ in June and $Q/Q_E = 0{\cdot}40$ during August for overcast days, we get $Q/Q_0 = 0{\cdot}66$ in June compared with 0·48 in August. From this we see that in order for the transmissions to have the observed value, either the optical thickness of the cloud or the surface albedo or both, must have changed from June to August. For example, referring to *Fig. 5*, h/L was about 16 in August, if the surface albedo was about 0·75 in agreement with the August 1955 measurement. Therefore, in June either the surface albedo was over 0·9 with the same cloud characteristics or else, as seems more likely, the cloud was optically thinner, in June. The possibility also remains that the cloud could have been optically thicker in June with a surface albedo of more than 0·8. But in that case the surface albedo in August must have been much less than 0·75.

It should also be pointed out that, with the presence of lakes and open leads in August, an albedo measurement over ice on T-3 might be too high to be representative of the whole surface area which reflects to the clouds.

Ice

This then was the picture of solar radiation versus cloud and surface albedo. If all the solar radiation were to go into melting ice, about 9 m of

ice would have melted from March to September. The observed melting was about zero according to ablation stake data on T-3 (GOLDSTEIN[4]). Estimates of an average of 1 m melt on the ice pack are accepted by CORTON.[5] In any case, much less than 9 m of ice is melted. Of course, some of the incoming energy is immediately reflected from the ice. If we again take the albedo to have been 0·75 in 1953, we can compute how much energy was absorbed by the ice. This is shown in *Fig. 7.*

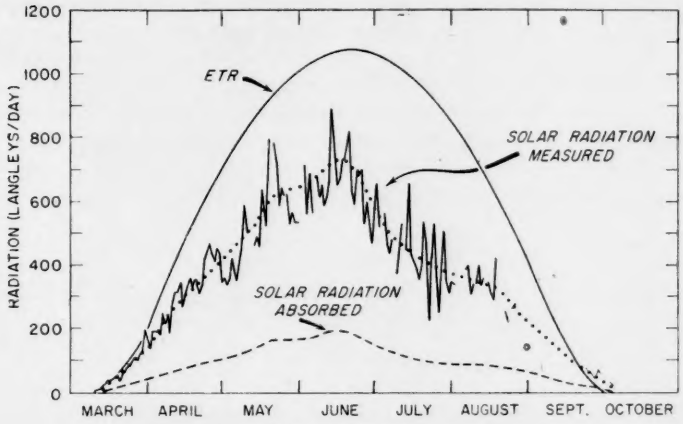


Fig. 7. Solar radiation absorbed by surface on T-3

Suppl. Part. I 1956

According to data obtained verbally from A. P. Crary of AFCRC, during 1953 about 1·6 ft of snow was present on the ice in April; this decreased to zero during the summer. According to him the density of snow was about 0·3; thus there would be 0·48 ft of equivalent water melted or about 14·7 g/cm² of water. Taking the heat of fusion to be about 80 cal/g, we get about 1150 cal/cm² required to produce the observed melting. In addition the temperature of the snow was raised from about −40°C to 0°C. This used up about 300 cal/cm². The total used to modify the snow was thus about 1450 cal/cm². This is about 8 ly/day.

Moreover, during 1953 very little ice was lost, if any; even allowing for larger melting in the lakes in T-3, only a small fraction of the absorbed energy could be accounted for.

Since the total melting of ice and snow can account for only a small portion of the solar energy absorbed, the heat must have been used or removed in other ways.

On the basis of a seasonal surface temperature change from about −30°C to 0°C, it is possible to compute the amount of heat added to a column of ice in order to raise its temperature. This heat added amounts to only about 0·3 of the solar heat absorbed by the ice.

Where the remainder of the solar energy was lost is speculative. Many heat transport mechanisms may be cited. Evaporation may use up a little. Transport to the warm continents may remove heat although it seems more likely to bring heat in. Finally, what seems most likely is turbulent heat transfer upward to the clouds. During summer the sky is nearly always

overcast. As a consequence there would be little or no radiative heat loss from the surface upward. Actually, on the Juneau ice field, heat was apparently added to the ice surface by long-wave radiation from the overcast sky, although in that program, advection was important in raising the cloud temperature (HUBLEY[6]). Nevertheless, a net long-wave radiation loss from the cloud top will go on continuously. This ought to cool the cloud until eventually if nothing else happens, an unstable lapse rate would be established under the cloud, especially since the ice surface temperature is fixed near 0°C by the melting process. This would result in turbulent heat transfer from the ice to the cloud; this heat added to the cloud would be lost by radiation from the cloud top, maintaining a more or less steady state. That an unstable lapse rate is actually established is evident from the radiosonde data on T-3. During the summer of 1953, on a great many days adiabatic and even super-adiabatic lapse rates were observed up to 300 ft and sometimes higher. This supports the possibility of an upward heat transfer.

ACKNOWLEDGMENTS

Thanks are due to A. P. Crary, N. Goldstein and other members of AFCRC who supplied the T-3 data and willingly discussed many features of the research.

REFERENCES

[1] MACDONALD, T. H. "Some Characteristics of the Eppley Pyrheliometer" *Mon. Weath. Rev.* Vol. 79, pp. 153–159, 1951

[2] LILJEQUIST, G. H. "Radiation, and Wind and Temperature Profiles over an Antarctic Snow Field—A Preliminary Note" *Proc. Toronto Met. Conf.*, pp. 78–87, 1953; published by Roy. Met. Soc., 1954 (see also: Norw.-Brit.-Swed. Antarctic Expedition, 1949–52, Scient. Results, Vol. 2, Part 1A, Oslo, 1956).

[3] FRITZ, S. "Illuminance and Luminance under Overcast Skies" *J. opt. Soc. Amer.* Vol. 45, pp. 820–825, 1955

[4] GOLDSTEIN, N. Private communication, 1956

[5] CORTON, E. L. "The Ice Budget of the Arctic Pack and its Application to Ice Forecasting" U.S. Navy Hydrographic Office, *Technical Report* No. 6, September 1954

[6] HUBLEY, R. C. "Micro-Meteorological Studies" *Prog. Rep. Juneau Ice Field Res. Proj., Alaska 1954*; Am. Geographic Soc., New York, February 1955 (Mimeographed Report)

Suppl.
Part. I
1956

LONG-WAVE RADIATION AND TURBULENT HEAT TRANSFER IN THE ANTARCTIC WINTER AND THE DEVELOPMENT OF SURFACE INVERSIONS

G. H. Liljequist
University of Stockholm, Sweden

In meteorological respect snow possesses properties which in a marked way affect the micrometeorological and also the macrometeorological conditions over extensive snowfields, such as, for instance, the ice-shelves and inland-ices of Antarctica. Snow is a very good reflector of short-wave radiation, i.e. solar radiation, but radiates itself practically as a black body, and last, but not least, snow is a bad conductor of heat. The conditions over an extensive snowfield are highly different at air temperatures below and above 0°C; at the former temperatures snow, and also the air in the surface layers, possess an extreme adaptability to varying meteorological conditions.

The conditions at air temperatures above freezing have been investigated over a number of glaciers in the Arctic and Sub-Arctic after the pioneering work of Sverdrup in 1934 on a Spitsbergen snowfield. The conditions at sub-freezing temperatures were studied by Sverdrup during the *Maud* expedition, 1918–25, to the North Polar Sea and a first quantitative idea of the energy components at work was then gained.

During the Norwegian-British-Swedish Antarctic Expedition, 1949–52, the meteorological program consisted of two principal investigations, besides the pure weather observations, viz. a study of the upper-air conditions and of the energy exchange of the ice-shelf at the base Maudheim (latitude 71° 03′S, longitude 10° 56′W Gr.). For the latter investigation we could scarcely have hoped for a better position of our base; it was situated upon a flat ice-shelf and about 5 km from the coastline proper. We could study the meteorological phenomena almost as ideally as when a physicist studies physical phenomena in a laboratory.

The investigation consisted of a study of the different energy components affecting the ice-shelf surface, viz.:

(1) direct solar radiation;
(2) global radiation;
(3) the albedo of the snow surface;
(4) the penetration of the visible radiation into the snow;
(5) long-wave radiation;
(6) radiation balance of the snow surface;
(7) the duration of sunshine;

(8) the turbulent heat transfer to the surface, obtained from a determination of the temperature- and wind-profiles in the surface layer;

(9) the heat conduction in the snow obtained from temperature and density measurements.

Each of these studies was a more or less independent investigation, but thanks to the excellent conditions the puzzle did fit surprisingly well.

The investigation is being published by Norsk Polarinstitutt, Oslo, in Vol. II, Part 1, of the Scientific Results of the expedition, and a first preliminary account was given at the Toronto meeting of 1953.

Wind- and temperature-profiles were determined in an observation mast, about 10 m high, *Fig. 1.* Along it were mounted six cup-anemometers and five platinum thermometers. The recording was done in one of the huts. The anemometers were constructed at the Norwegian Meteorological Office and were designed to stand overall Antarctic conditions well. The number of contacts were counted by electromagnetic counters, which were automatically photographed once every hour, thus giving the mean wind speed in the 1-hr interval between exposures. The temperature recording built upon the principle of an unbalanced Wheatstone bridge. This was also the case with the temperature-reading for checking the recording. This outfit had been supplied by the Cambridge Instrument Company, Cambridge, England. In order to eliminate the radiation-error the platinum thermometers were mounted in between two pairs of aluminium shields with upper and lower surfaces polished. The ventilation was thus done by the wind. Measurements and considerations show that there is a radiation-error in the summer, but in spite of that the temperature gradient is accurately given by the two uppermost thermometers, viz. at 5 and 10 m. In the winter half-year the radiation-error can, however, be considered nil.

We will limit ourselves to giving a rough outline of the conditions in connection with the development and maintenance of surface inversions in the winter.

With a clear sky and light to moderate wind, marked surface inversions developed in the winter; a temperature difference of 10–20°C between top and foot of the 10-m high mast was then usual, *Figs. 2* and *3.* Marked temperature fluctuations occurred, generally brought about by a fluctuation in level of the different strata of air. These temperature fluctuations decreased in magnitude towards the surface except in the case of a marked thermocline when maximum fluctuation occurred at that level. Sometimes also a rolling up of the inversion by gusty or squally winds gave rise to a marked temperature fluctuation. On some of these occasions the rolling up took place at the station, on others some distance upwind from it. In the latter case a lag in the occurrence of the fluctuation was apparent in our temperature diagrams, the change occurring later at the lower levels, having a smaller wind speed, than higher up, see *Fig. 2* at 05^h–06^h. There was also a typical similarity in wind- and temperature-variations.

As to the temperature-profiles it was apparent that they were often linear or in part linear, though "irregular" types were also very common, *Fig. 3.* Sometimes a marked thermocline separated layers with less stability above and below.

uppl.
art.I
956

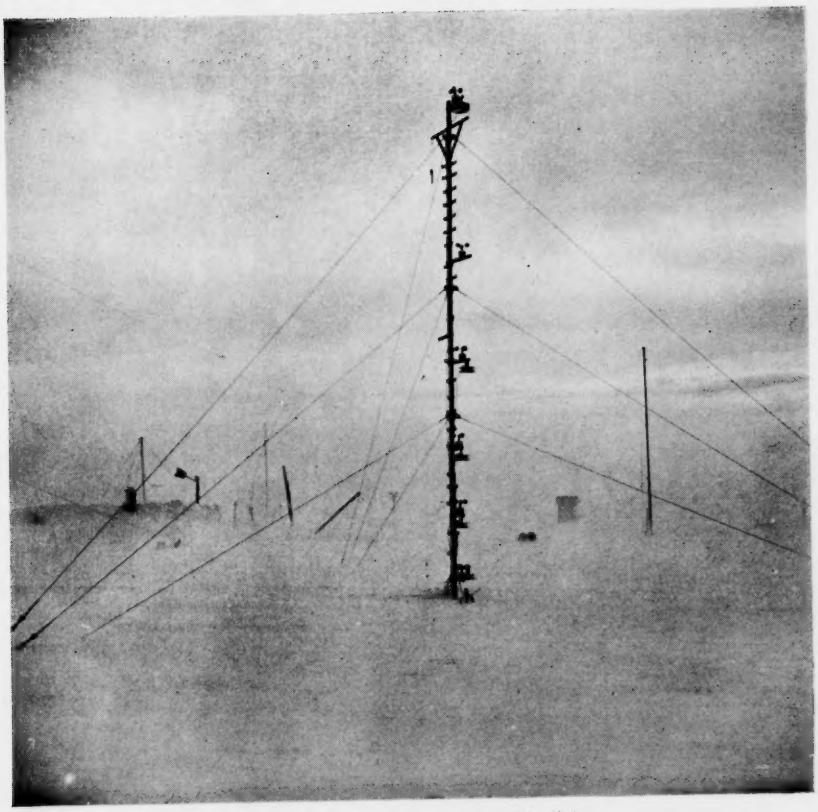


Fig. 1. The meteorological mast at Maudheim

(*Facing page* 168)

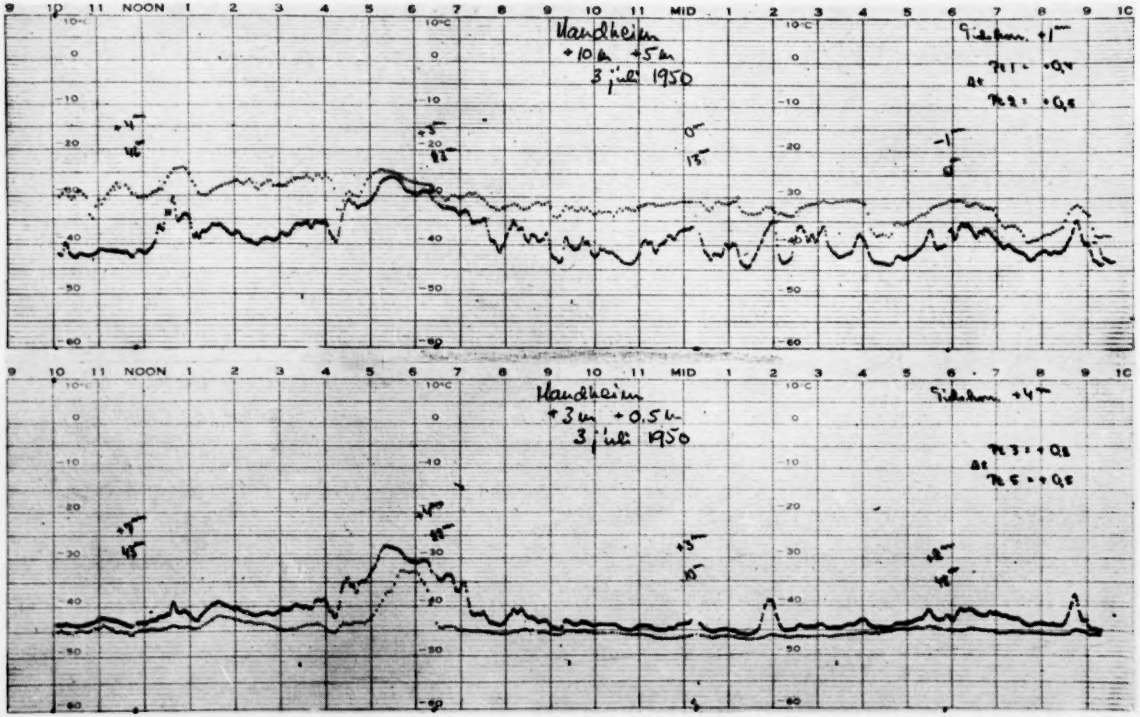


Fig. 2. Examples of temperature records from a fine-weather situation in the winter. The heights, which are entered in the diagrams, were only used as reference at Maudheim, compare Fig. 3. Further, "Mid" and "Noon" of the diagrams should be interchanged

(Facing page 164)

Sup
Par
195

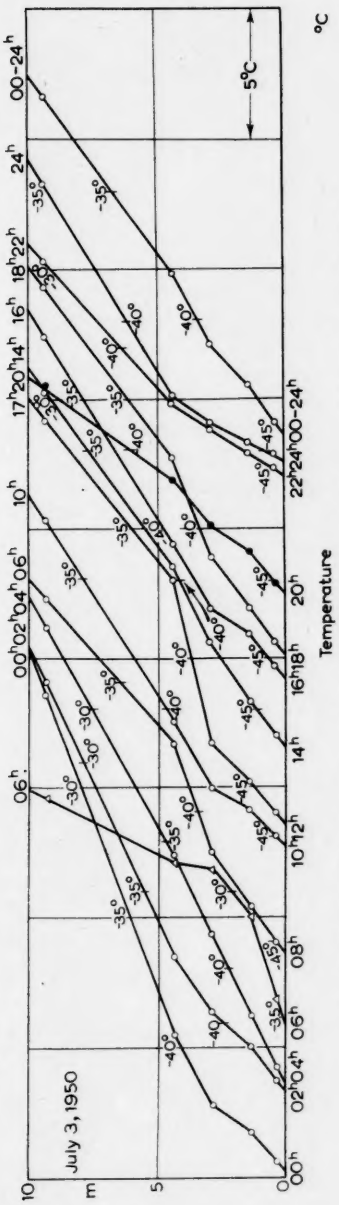


Fig. 3. Example of temperature-profiles from the same day as the diagrams of Fig. 2, viz. 3 July 1950. The temperatures are hourly means centred at even hours

Suppl.
Part. I
1956

There is a marked difference between the physical factors governing the inversions developing over a snowfield at air temperatures well above and well below 0°C. In the former case the foot of the temperature-height curve is fixed at 0°C independent of the variations of the air temperature. In the latter case the temperature at the surface is free to vary, as imposed by the meteorological conditions. As previously pointed out the snow-surface and the immediately overlying air-layers then possess an extreme adaptability with regard to temperature variations. This type of inversion is built up owing to a negative radiation balance at the surface or, rather, owing to the fact that heat conduction from the layers below the surface cannot cover the energy losses due to radiation. This type of inversion is thus a sort of compensation phenomenon.

If the radiation balance becomes positive, i.e. if the net radiation is incoming, this inversion will disappear. In the winter with a clouding-over to a dense overcast, the net long-wave radiation will, to begin with, be incoming instead of outgoing, as the cloud-base is situated well above the principal portion of the inversion. The surface inversion will then disappear, until for all practical purposes cloud-base and snow-surface have both attained the same temperature, and the net radiation is nil. In this connection we should remember that heat conduction in the snow will gradually

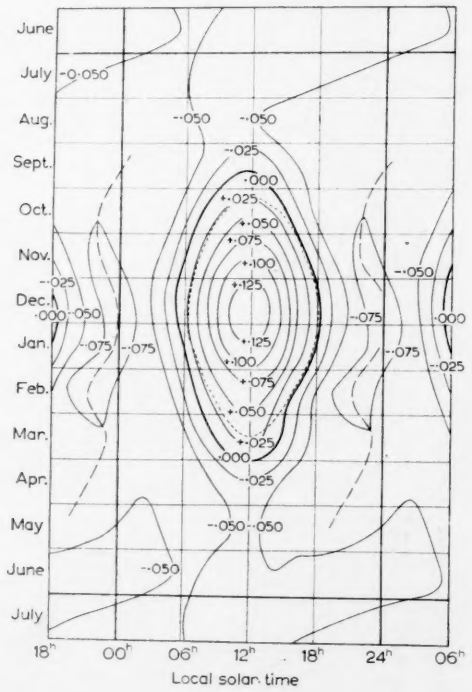


Fig. 4. Isopleth diagram showing the radiation balance of the snow surface in fine-weather situations, i.e. clear sky and light to moderate wind. Zero-balance with actual inversion-conditions has been marked with a thick line, with no inversion with a dotted line. Energy-unit: ly/min

become insignificant, and further, with the inversion gone, the weak turbulence of the air over the flat and smooth ice-shelf cannot maintain an indifferent stratification up to the level of the cloud-base. Radiation will therefore be the deciding factor.

In the summer season the radiation balance will become positive in day-time owing to the short-wave radiation flux from the sun (see *Fig. 4*), where zero-balance has been drawn with a thick line for the normal inversion conditions obtaining with a clear sky. As is seen, the balance is positive in day-time for around 12 hr at midsummer when the sky is clear. With no inversion the net long-wave radiation is greater than with inversion, compare below. In *Fig. 4* we have therefore also drawn radiation balance-zero also for non-inversion conditions, the dotted line. Within the area

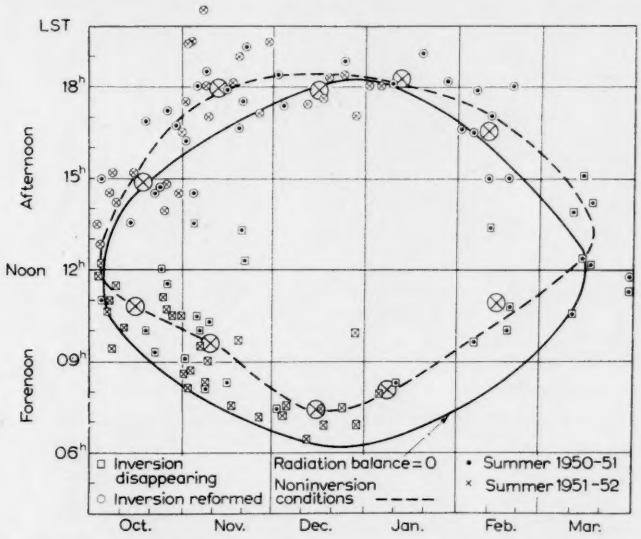


Fig. 5. Times of formation and destruction of the inversion in the surface layer due to solar radiation, and comparison with the times of radiation balance equal to zero with no inversion, compare Fig. 4

determined by this latter closed curve, non-inversion conditions can thus be maintained. In *Fig. 5* we have compared the times of the formation and disappearance of surface inversions (as given by the thermometers at 5 and 10 m) on more or less clear days with this latter zero-balance curve. Considering the lag, which must occur in the building up and the destruction of inversions, the agreement between the two curves must be considered good.

In the height of summer with strong solar radiation and with light wind the temperature-profile is not correctly given by the platinum thermometers near the surface owing to the radiation-error affecting them. In illustration of the effect of *incoming* net radiation upon the inversions we have therefore in *Fig. 6* given some profiles occurring in connection with a clouding-over in the winter. The *incoming* (long-wave) radiation was on this occasion at 18–19^h around 0·040 ly/min. As is seen from this example inversion persists in the layers above approximately 1·5 m, while a lapse is found between

Suppl. Part. I 1956

1·5 m and the surface. In other words, the layers around 1·5 m act as a heat sink with turbulent heat transfer above, and free convection below. The disappearance of the inversion and the gradual heating of the surface layers appear clearly from this diagram. Similar conditions will occur during the summer season in the morning or forenoon.

Surface inversion (temperature-profile) and wind-profile affect each other mutually, and from them the turbulent heat transfer is determined. The surface inversion also affects the net long-wave radiation from the surface. With a fine-weather situation in the winter a balance develops between the

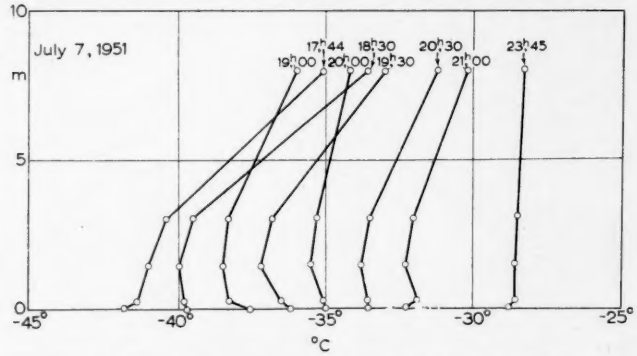


Fig. 6. Disappearance of the surface inversion in connection with a clouding-over in the winter, giving rise to incoming net long-wave radiation

energy-loss of the surface, i.e. net long-wave radiation, and the energy-income, i.e. turbulent heat transfer in the air and heat conduction in the snow—such are generally the directions of the energy fluxes with clear sky in winter-time.

The heat conduction in the snow can easily be computed with the help of temperature measurements at a number of points in the uppermost 10–15 cm of the snow, plus a determination of the snow-density. In fine-weather situation in the winter 0·020 ly/min can be considered as a good average.

Previous measurements of net long-wave radiation in the polar regions—notably the series from the *Maud* expedition to the North Polar Sea, 1918–25, and from Fairbanks, Alaska—have shown how the net radiation decreases with decreasing temperature (see *Fig. 7*), compiled from WEXLER,[7] where also the Maudheim data have been entered. This decrease has been ascribed to the development of surface inversions.

At Maudheim the net radiation from the surface took on values around 0·100 ly/min, when an inversion was lacking or almost lacking, e.g. in connection with the dying down of a gale with the sky clear, or immediately after a clearing-up after a dense overcast. With a fine-weather situation, i.e. clear sky and light to moderate wind, the net long-wave radiation settled at values around 0·050 ly/min, when an inversion had developed.

We undertook one radiosonde ascent a day at Maudheim. In order to find out how surface inversions affected long-wave radiation, we determined from these ascents the "tropospheric temperature maximum", T_x, i.e. the highest temperature in the troposphere found on top of the inversion, on an

average at 500–1000 m in the winter. From this level the temperature will decrease with height. We will as parameters use this T_x and also the magnitude of the surface inversion ΔT, i.e. $\Delta T = T_x - T_o$, where T_o is the temperature at pyrgeometer level. The conditions in the upper atmosphere are thus represented with T_x, the inversion by ΔT. Here should be added that on an average 40–45 per cent of ΔT was concentrated to the lowest 10 m.

In this connection it should also be mentioned that the directly measured values of the net long-wave radiation generally have no meteorological significance. The instrument, in our case an Ångström pyrgeometer situated at about 1·5 m, gives the net radiation at the level of the instrument. The measured values must be reduced to the surface, which can be done by

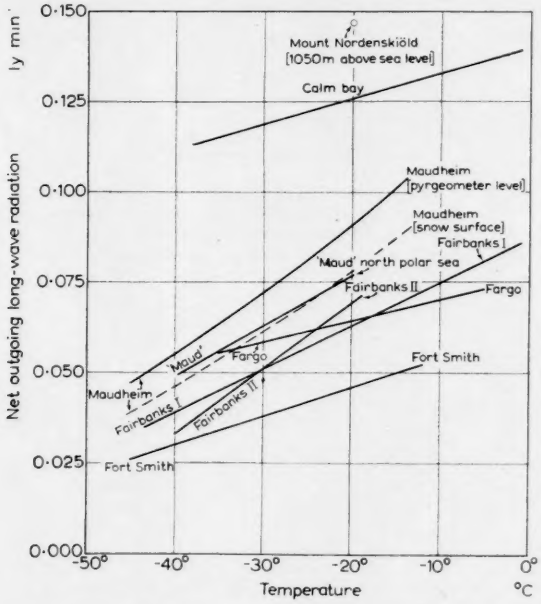


Fig. 7. Average net long-wave radiation as a function of the "surface" temperature (or rather screen temperature)

also facing the instrument downwards and taking the difference of the two measured values. One can also place a thermometer on the snow to obtain the approximate surface temperature and then compute a correction to be applied. In our case these two methods gave the same result. With a surface inversion this correction was of the order of magnitude 0·010 ly/min, i.e. it cannot be neglected. We found that the correction was a function of the measured net radiation with the pyrgeometer facing upwards.

With a surface inversion developing, the long-wave radiation from the atmosphere must decrease; the atmosphere grows colder. Let us, to begin with, find out how the atmospheric long-wave radiation R_A is affected by the inversion. We will use our two parameters mentioned above, viz. T_x and ΔT. *Fig. 8* will show this relationship given by group-means in order to let the variations appear more clearly.

Suppl.
Part. I
1956

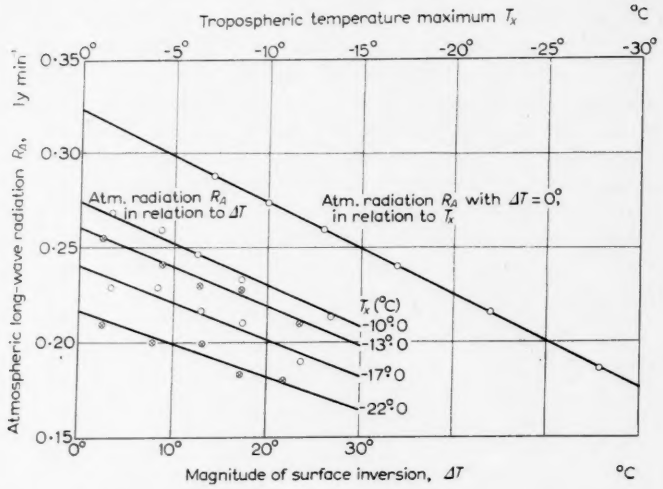


Fig. 8. The atmospheric long-wave radiation R_A decreases by about 0·8 per cent per 1°C increase in the magnitude ΔT of the surface inversion. R_A depends further upon the temperature conditions in the atmosphere above the inversion, here represented by the temperature T_x on top of the inversion. The data in the figure have been obtained by grouping our observations

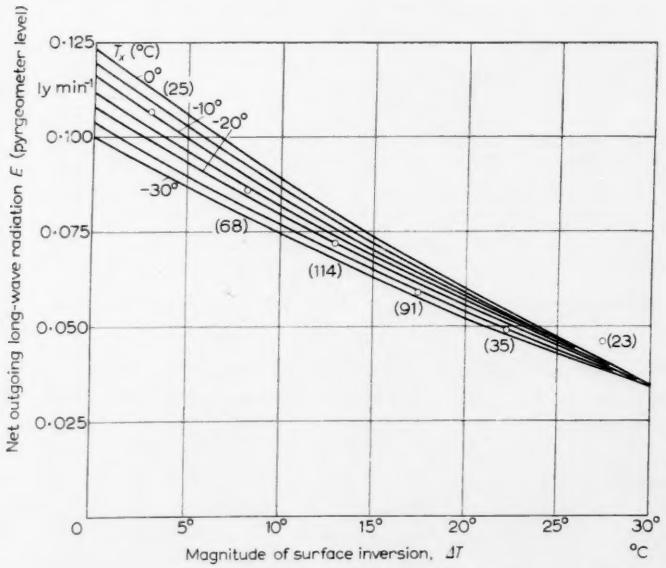


Fig. 9. The net long-wave radiation E (at pyrgeometer level and pyrgeometer temperature) as a function of the magnitude ΔT of the surface inversion and of the temperature T_x on top of the inversion. With a well-developed inversion the influence of T_x upon E decreases, and the latter may for practical purposes here be considered as a function of ΔT alone. Note: This relationship has been obtained not from the values of R_A given in Fig. 8, but from relations where some attention has also been paid to humidity. The differences are, however, immaterial

On an average the atmospheric long-wave radiation decreases by 0·8 per cent per 1°C increase in magnitude ΔT of the inversion. As is seen also the affect of T_x upon the atmospheric radiation is clearly given.

If the atmospheric long-wave radiation R_A is subtracted from the black-body radiation at temperature T_0, we obtain the net outgoing long-wave radiation E, that would be measured with the pyrgeometer at temperature T_0, *Fig. 9.* As is seen from this diagram E decreases from values between 0·100–0·125 ly/min with no inversion to values around 0·050 ly/min with

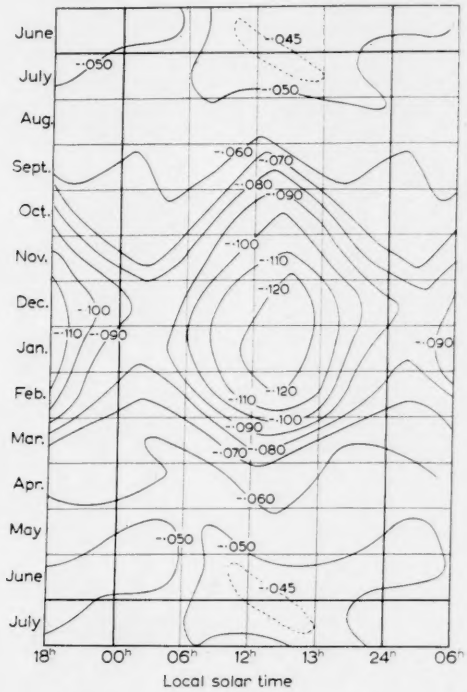


Fig. 10. Net long-wave radiation Es from the snow-surface in fine weather situations. Energy-unit: ly/min

a well-developed inversion, i.e. for $\Delta T = 20$–25°C. It is also apparent from the figure that the net radiation becomes less and less affected by the temperature conditions in the atmosphere above the inversion, i.e. by T_x, as the inversion grows in magnitude.

From these average values of net radiation E at pyrgeometer temperature T_0, we have still to subtract a correction in order to obtain the net long-wave radiation E_S from the snow-surface, compare above.

With radiosonde data and with temperature data from screen level (= pyrgeometer level) we are thus able to compute the average net long-wave radiation with a clear sky over the year. In other words, we are able to pay attention to the change in T_x and in the magnitude of the inversion which occur from season to season and from day to night, *Fig. 10.* The net

radiation given here is valid for clear sky and light to moderate wind, i.e. for fine-weather situations. With increasing wind the inversions are, however, rolled up by turbulence, and the net radiation increases, in the winter season from values around 0·050 to 0·100 ly/min.

The rolling-up of the inversions by turbulence was very characteristic. We have prepared *Fig. 11* to show this. From the two uppermost platinum thermometers, with a difference in level of 5 m, we have for each month in the winter half-year determined the highest observed temperature gradient for different wind-speeds within 1 m/sec intervals. A curve enclosing these points has been drawn to show the maximum observed temperature gradient for different wind speeds. As is seen, marked inversions can develop for wind speeds less than 8–9 m/sec at 10 m, while for higher winds the inversions, which do occur, are inconsiderable. This limit is remarkable in another

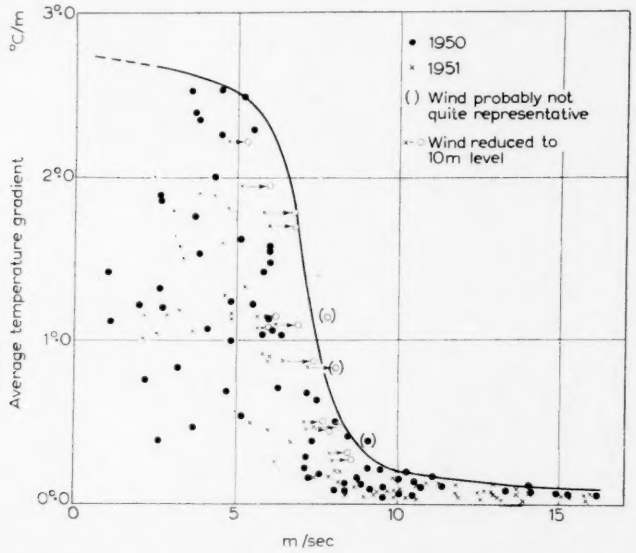


Fig. 11. The curve shows maximum observed temperature gradient (between approximately 5 and 10 m) for different wind speeds. The direct observations refer to maximum gradient observed each month during the winter half-year and within 1 m/sec intervals. The diagram shows the partial destruction of the inversion with winds higher than 8–9 m/sec

sense too. With loose snow available on the surface, the snow started drifting also at 8–9 m/sec. With Nikuradse's test for aerodynamically smooth and fully-rough flow we obtained further that the flow starts to become fully-rough also at 8–9 m/sec, i.e. fully-rough for winds higher than 8–9 m/sec and transitional to smooth flow for less wind speed.

With a dense overcast in the winter and with winds of gale force, there was generally indifferent stratification of the air and the wind-profile in the surface layer was, according to the logarithmic wind law, viz.

$$u = \frac{u_*}{\varkappa} \ln \frac{z+z_0}{z_0},$$

where $\varkappa$ is the Kàrmàn constant (= 0·40), u_* the so-called friction velocity, and z_0 the roughness parameter. In our case we found z_0 equal to 0·10 mm, except for winds higher than 13 m/sec, where it increased with the wind speed, probably owing to the friction of the drifting snow against the surface.

When a surface inversion develops the wind-profile is affected by the changing stability of the air. We found that, with increasing stability, the logarithmic profile changed, to become more and more linear in shape. We also found that the profile was well represented by splitting the wind into two components, one, u_a, being logarithmic and the other, u_l, linear and proportional to the height above the surface. With neutral stability the linear component is zero, but with increasing stability of the air the linear component increases to become dominating in situations with a well-developed inversion, *Fig. 12.*

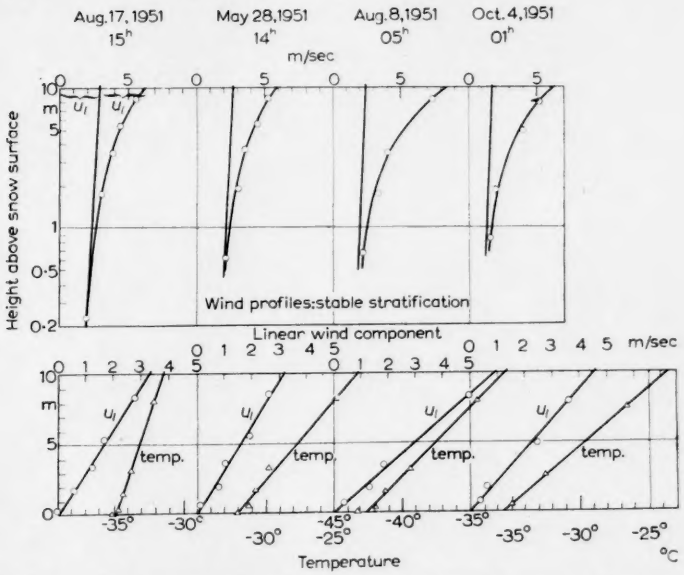


Fig. 12. Wind- and temperature-profiles. The wind-profile can be well represented by splitting the wind speed into two components, one being logarithmic, u_a and the other linear, u_l

The wind-profile can be described by defining a number $k = u_l/u_{10}$ (valid for a reference level, in our case 10 m). The relation between k and the stability S of the air is given in *Fig. 13.* For that purpose we have chosen more or less linear temperature-profiles, where S can be attributed one definite value. For further details reference is made to LILJEQUIST[1] or to the full expedition report, viz. Vol. II, Part 1C.

It is apparent (see, for instance, *Fig. 12*) that the wind-profile close to the surface will differ very little from the logarithmic wind component, referred to above. With a profile given against log z, as in *Fig. 12*, the wind-profile will approach the logarithmic component asymptotically for decreasing values of z. In other words, the surface friction exercised by the wind is given by the logarithmic wind component. The shearing stress τ in the low layer

may be taken to be constant with height and equal to the surface friction ($= \tau_0$), steady conditions being supposed to occur. From the equation

$$\tau = \rho K_M \frac{\delta u}{\delta z}$$

the eddy viscosity K_M can thus be computed, as we do know $\tau = \tau_0$ ($= \rho u_*^2$) and $\delta u/\delta z$ from the wind-profile. K_M is then found to be proportional to the wind speed at 10 m, and with a well-developed inversion it increases only slowly with height, except near the surface.

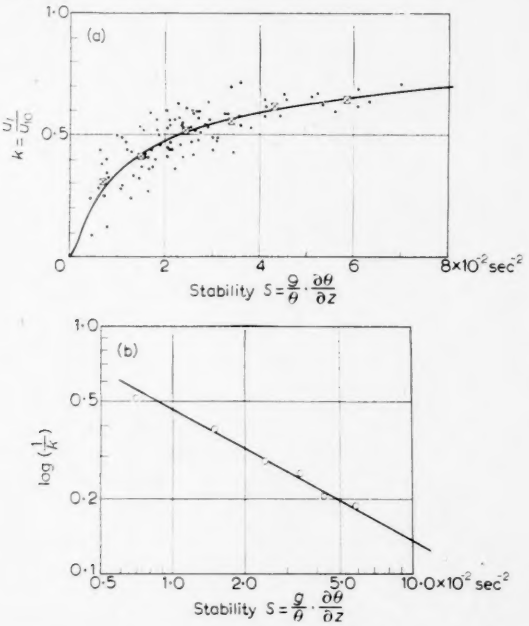


Fig. 13. Relationship between linearity parameter k ($= u_l/u_{10}$, *valid at 10 m*) *and stability* S *of the air*

Theoretically the eddy convectivity K_H should be equal to K_M, but in practice different authors are of different opinion on this point. Let us, however, to begin with suppose that they *are* equal and let us see what results we then obtain. These results can then be tested with direct measurements. The turbulent heat transfer Q can be written

$$Q = c_p \rho K_H \frac{\delta\theta}{\delta z},$$

where we suppose $K_H = K_M = K$.

Here θ is the potential temperature, which in our case well can be taken as equal to the actual temperature.

But $S = \frac{g}{\theta}\frac{\delta\theta}{\delta z}$ and further, considering one and the same surface pressure p_o (for Maudheim = 985 mb) we have $\rho/\rho_o = \theta_o/\theta$ where θ is the temperature (more strictly the potential temperature) at a middle level, in our case at 5 m, and $\theta_o = 273°\text{K}$.

We then obtain

$$Q = \frac{c_p\, \theta_o\, \rho_o}{g}\, S{\cdot}K \text{ (ly/sec)}.$$

Except for inconsiderable inversions, K will vary only slowly with height, compare above. Let us here consider K valid for the 5-m level, for which level perhaps our stability-values S also best may be referred. As K is proportional to the wind speed at 10 m, this is also the case with Q. For wind speed 1 m/sec and with Q given in ly/min (gcal cm^{-2} min^{-1}) we then obtain the heat transfer Q (here called Q_1), *Fig. 14*. It is seen that, excepting

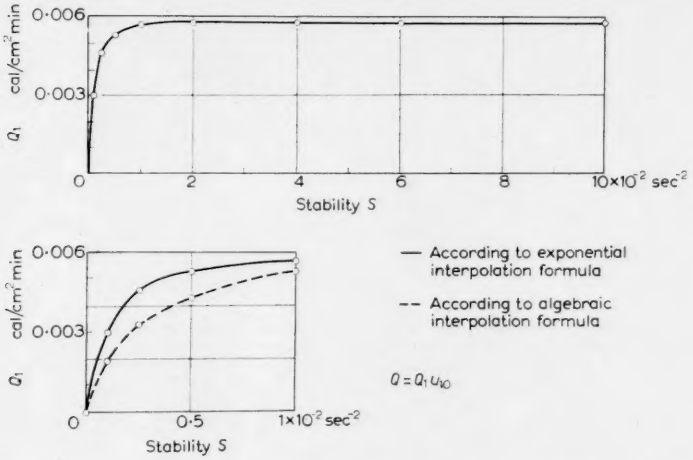


Fig. 14. Turbulent heat transfer, with wind speed 1 m/sec, at 10 m, as a function of the stability S of the air (S = 0·01 sec^{-2} corresponds approximately to a temperature gradient of 2·5°C per 10 m). With a well-developed inversion the heat transfer is independent of the stability (temperature gradient) and becomes simply proportional to the wind speed

inconsiderable inversions with gradients less than about 2·5°C per 10 m ($S \simeq 0{\cdot}01$ sec^{-2}), Q_1 is constant and equal to 0·0058 ly/min. With wind speed u_{10}, given in m/sec, the heat transfer is thus

$$Q = 0{\cdot}0058 \cdot u_{10} \text{ (ly/min)}.$$

During the *Maud* expedition, 1918–25, Sverdrup (using other considerations) also found proportionality between turbulent heat transfer Q and wind speed. He obtained a factor 0·0044 (against our 0·0058). The difference between these two values is not great.

The validity of our relation can be tested in fine-weather situations in the winter, when we possess several measurements of the net long-wave radiation E_S. For periods of 3–12 hr we can compute the average heat conduction σ in the snow, and we can also obtain a fairly good mean value of E_S for the considered period, from 2 to 5 measurements. The difference $E_S - \sigma$ should in steady conditions be equal to Q, if we neglect sublimation or evaporation. In our case sublimation did practically not occur in our 20 situations chosen, and evaporation should be immaterial at the low temperatures in the winter.

Suppl.
Part. I
1956

In *Fig. 15* we have entered these differences against the mean wind speed u_{10} for respective periods, and we have also drawn the line $Q = 0{\cdot}0058\, u_{10}$ (compare above). The points are fairly well grouped along this line, and we may therefore consider our derivation of the turbulent heat transfer as verified.

With a more or less constant *average* heat conduction in the snow, we should expect the net long-wave radiation E_S in fine-weather situations in the winter to be a linear function of the wind speed. This is verified by the

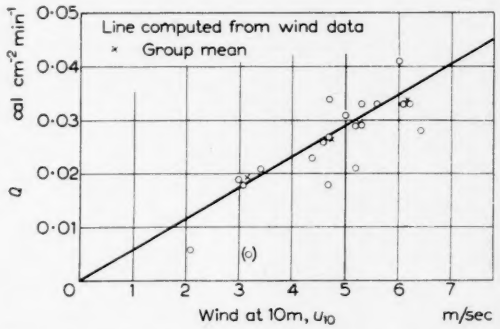


Fig. 15. Test of the relationship between wind speed and turbulent heat transfer Q. The straight line has been obtained from wind and temperature profiles, the points from measurements of net long-wave radiation, heat conduction in the snow and wind speed at 10 m

Maudheim measurements, and we find also from this relation the turbulent heat transfer to be about $0{\cdot}0060\, u_{10}$.

The similarity in wind and temperature variations which are observed in the polar winter in fine-weather situations is a consequence of the interplay between net long-wave radiation, turbulent heat transfer and heat conduction. I only mention the fact that generally maximum temperature and maximum wind speed are observed at or a little after midnight, while minimum values are found on an average at noon or in the afternoon.

The energy-exchange at the snow-surface in connection with well-developed surface inversions is small. Let $u_{10} = 5$ m/sec represent average wind speed in fine-weather situations. The order of magnitudes of the energy fluxes to be considered are then:

Turbulent heat transfer	0·030 ly/min
Heat conduction in the snow	0·020 ly/min
Net outgoing long-wave radiation	0·050 ly/min

In the winter the snow is protected from heavy energy-losses by the development of a coating of cold air (an inversion), which makes the radiation-losses small and which besides sets up a turbulent flux of heat from the air to the surface. The net energy-loss of the snow-surface *must* become small so that the snow—with its low heat-conductivity—really can meet with these losses. In other words, the development of the marked surface inversions in polar regions is mainly due to the low heat-conductivity of snow.

References

[1] LILJEQUIST, G. H. "Radiation, and Wind and Temperature Profiles over an Antarctic Snow-field—a Preliminary Note" *Proc. Toronto Met. Conf.*, 1953; Roy. Met. Soc., London, 1954

[2] "Energy Exchange of an Antarctic Snow-field" *Scientific Results of the Norwegian-British-Swedish Antarctic Expedition, 1949–52* Vol. II, Part 1; published by Norsk Polarinstitutt, Oslo, 1956–57

[3] SUTTON, O. G. *Micrometeorology. A Study of Physical Processes in the Lowest Layers of the Earth's Atmosphere;* McGraw-Hill, New York, Toronto, London, 1953

[4] SVERDRUP, H. U. Meteorology, Part I, Discussion. Norwegian North Polar Expedition with the *Maud*, 1918–25. *Scientific Results.* Vol. 2, Bergen, 1933

[5] ——. "The Ablation on Isachsen's Plateau and on the Fourteenth of July Glacier in Relation to Radiation and Meteorological Conditions" *Geogr. Ann.* Vol. 17 (3–4), 1935

[6] ——. "The Eddy Conductivity of the Air over a Smooth Snow-field" Results of the Norwegian-Swedish Spitsbergen Expedition in 1934. *Geofys. Publ.* Vol. 11 (7), 1936

[7] WEXLER, H. "Observations of Nocturnal Radiation at Fairbanks, Alaska, and Fargo, North Dakota" *Mon. Weath. Rev. Suppl.* No. 46, 1941

Suppl.
Part.I
1956

ON THE "ARCTIC WHITEOUT"

SIGMUND FRITZ
U.S. Weather Bureau, Washington, D.C.

HEDINE[1] described the dangerous "Arctic whiteout" in these words: "A peculiar condition affecting visibility occurs in north-western Alaska during late winter and early spring. The descriptive term "Arctic whiteout" is applied to it as it camouflages all visible land features, blending earth and sky so that the horizon and all landmarks are indistinguishable.

"A moderate snow of late winter can obliterate all vegetation and topographical relief; when, in addition, there is an overcast of cirrostratus and/or altostratus clouds, no shadows are cast; the eye cannot determine what landmarks may or may not be visible. It is impossible to tell by visual check from a plane whether or not the aircraft is upside down. Hills and mountains are blotted out."

COURT,[2] who prefers the term "milky weather" to "whiteout", quotes Arctic veterans who tell of "stumbling and falling because irregularities in the generally level snow surface are completely invisible".

Further descriptions of "whiteout" conditions appear in STEFANSSON's publications.[3, 4] During these conditions black objects would be visible at large distances.

To explain all the details of the "whiteout" would require an involved psycho-physical investigation. For example, the threshold contrast between an object and its background may be different during "whiteout" conditions than it is in the laboratory or over ordinary terrain. However, in order to derive some relations we shall assume that the threshold contrast for invisibility is nearly zero, as it is in fact under ordinary conditions in middle latitudes.

To delimit the problem, let us assume that under "whiteout" conditions the horizontal ground at the horizon is indistinguishable against the horizon sky, and further that a vertical plane surface, with the same albedo as the ground, is also invisible against the horizon sky. These criteria seem appropriate to explain some of the properties described earlier by Hedine and Court, namely, that the horizon as well as hills and vertical protuberances are invisible against the sky.

The basic requirements can be illustrated with the aid of *Fig. 1.* In this figure a vertical surface is shown on the horizontal ground. The amount of light which falls on the vertical surface will depend on the *distribution* of the luminance* of the sky and of the ground. The amount of light which reaches the observer from the vertical surface will further depend on the albedo of that surface. We are requiring as part of the model of the "whiteout" that

* Luminance, or brightness, is the amount of light which reaches an observer from unit projected area of sky, per unit solid angle, per unit time.

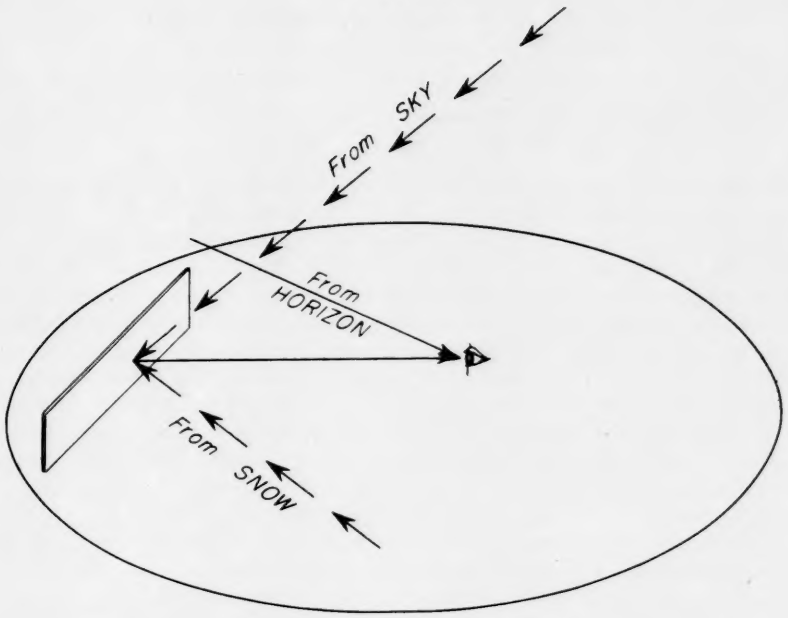


Fig. 1. The amount of light which reaches an observer. Further details given in text

the light coming to the observer from the vertical surface should be the same as that coming to him from the horizon sky. Similar statements can also be made about the horizontal ground.

MIDDLETON[5] has discussed the luminance of both the ground and vertical grey surface under overcast sky conditions. He assumed that the luminance of the overcast sky, $B_s(\theta)$, is given by

$$B_s(\theta) = B_h\,(1+n\cos\theta). \tag{1}$$

Here, θ is the angle measured from the zenith as zero and B_h is the luminance of the horizon sky. By assuming further that the ground was a perfectly diffuse reflector with albedo R, Middleton computed the luminance of a white vertical surface.

According to observations which were available, in addition to some which Middleton made himself, all made over snow-free terrain, it was found that $n = 2$ gave a good representation of the sky brightness distribution; that is, it was found that the zenith sky is about three times brighter than the horizon sky during overcast conditions.

It must be emphasized that these measurements were made over snow-free terrain. However, on the basis of theoretical calculations (FRITZ[6]), it can be shown that n decreases as R, the ground albedo, increases. If this is so, the brightness of the overcast sky becomes much more uniform over snow cover, and any conclusions regarding the visibility of vertical white objects would have to be modified in an important degree.

Suppl.
Part.I
1956

To test this further, we have assumed the following relations:

For the overcast sky, the luminance is again given by

$$B_s(\theta) = B_h\,(1+n\cos\theta). \quad 0 \leqslant \theta \leqslant \pi/2. \tag{1}$$

For the ground, the luminance is given by

$$B_g(\theta) = B_h\,(1+p\cos\theta). \quad \pi/2 \leqslant \theta \leqslant \pi. \tag{2}$$

At the horizon, $\theta = \pi/2$. For the horizon, therefore, the horizon sky automatically has the same luminance as the horizon ground, for $B_s(\theta) = B_h = B_g(\theta)$ at $\theta = \pi/2$. Actually in the computations the usual threshold relation between ground and sky was also considered, although this has little effect on the numerical results. It remains to evaluate the constants n and p. Two equations can be formed. One equation is formed from the fact that the energy reflected from the horizontal ground depends on the sky brightness distribution (Eq. 1) and the ground albedo; moreover, the upward directed energy can also be computed from Eq. (2). Another equation can be formed by computing the luminance of a perfectly diffuse *vertical* surface whose albedo is the same as the ground albedo. These two equations thus obtained can be solved for n and p in terms of the ground albedo R.

The results for the sky luminance distribution are given in *Fig. 2*. This figure shows the luminance of the sky, $B_s(\theta)$, as a function of the angle, θ, from the zenith. The results have been normalized to unity at the zenith. Each curve is for a different value of snow reflectivity, R. The curve for $n = 2$, which corresponds to available measurements of sky luminance over snow-free terrain, is shown as the heavy dashed curve. We also see that when the albedo of the ground or snow surface is close to 1, the sky luminance distribution is just about uniform for all values of θ. The sky luminance distribution remains fairly uniform for $R = 0{\cdot}9$. But the sky becomes less uniform, approaching the luminance distribution for ordinary snow-free conditions, when $R = 0{\cdot}65$, *provided that* "whiteout" conditions are assumed.

It should be noted that the mathematical model used to define the "whiteout" in this paper will permit a sky brightness distribution to be computed for all values of surface albedo, R; also with this mathematical model, the sky luminance becomes less and less uniform as the surface albedo decreases.

Since from experience the "whiteout" cannot occur for all values of surface albedo, it is necessary to limit the problem further by introducing a third condition in addition to the two equations mentioned earlier. Qualitatively, we may conclude that the "whiteout" does not occur if the surface albedo is less than 0·6, for the following reasons. First, when the sky luminance is so non-uniform that the zenith is three times brighter than the horizon, it seems likely that a pilot could determine the vertical direction by noting the brightest area of the sky under such conditions; this is the case when $R = 0{\cdot}65$ or $n = 2$. Secondly, as mentioned earlier, when the surface albedo is close to zero, observation shows that the zenith is about three times brighter than the horizon. Raising the surface albedo to 0·6 will probably make the sky at least a little more uniform; therefore the curves for $R \leqslant 0{\cdot}6$ in *Fig. 2* would be inconsistent with observed sky luminance distributions.

Although we can decide that 0·6 is too low an albedo to permit the "whiteout", it is not possible, with the theory as presented, to state a lower

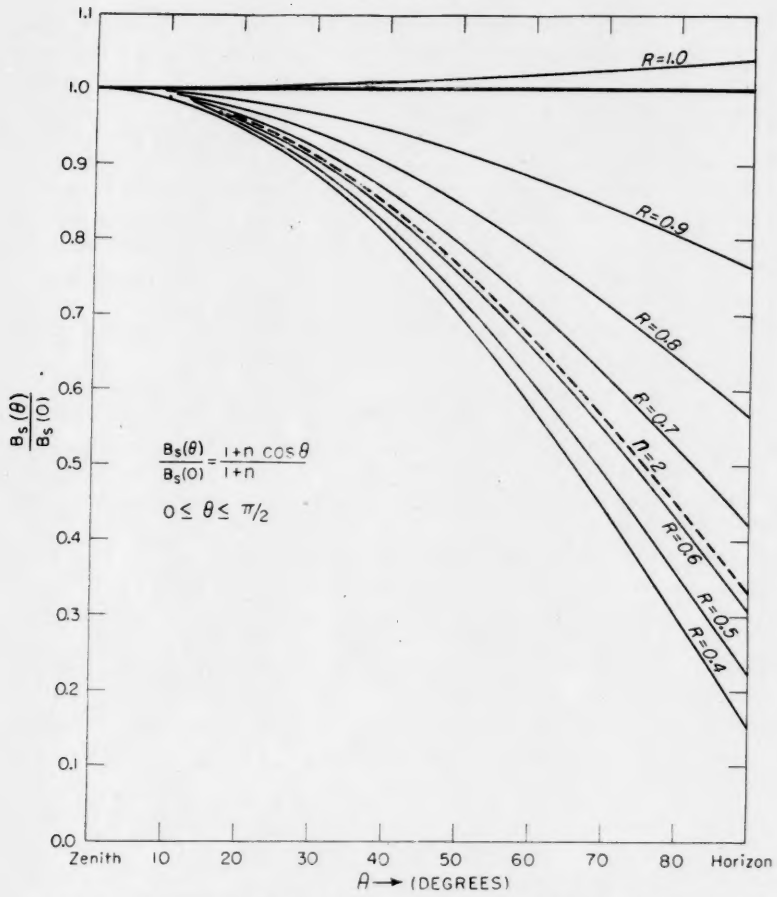


Fig. 2. The luminance of the sky, $B_s(\theta)$, as a function of the angle, θ, from the zenith

limit for surface albedo under "whiteout" conditions. However, since 0·6 is already too low an albedo, if we could observe the horizon in "light" for which the surface albedo was 0·6 or less, the horizon would be discernible even though it could not be seen by the unaided human eye. If it were possible, for example, to observe the horizon with radiation at 10μ, a contrast between cloud and surface would exist provided the cloud temperature is sufficiently different from the surface temperature. With the small amount of water vapor present in polar regions, radiation at 10μ might penetrate relatively unmodified to the observer from the horizon. Perhaps some of the more recent infrared detecting devices might be adapted for this purpose, even at wavelengths near 2μ if the snow albedo is low enough.

An interesting additional point is concerned with the parameter p which appeared in Eq. (2). This parameter determines the brightness distribution of the ground, and although small it is not equal to zero. The computations

indicate that during "whiteout" conditions the nadir is somewhat brighter than the horizon, the difference increasing as the ground albedo decreases. It should be pointed out that MIDDLETON and MUNGALL[7] have found that snow is generally not a perfectly diffuse reflector. Therefore, a non-zero value for p is not entirely unexpected since $p = 0$ corresponds to a perfectly diffuse reflecting snow surface.

Finally, it would have been highly desirable to compare these conclusions about sky luminance distribution with observations. But unfortunately no suitable observations have so far been made.* Since a good understanding of the "whiteout" will depend on the acquisition of suitable measurements, plans are well advanced to measure the sky luminance distribution in the Antarctic during the International Geophysical Year. Simultaneous surface albedo measurements in visible light are planned also.

REFERENCES

[1] HEDINE, L. J. C. "The Arctic Whiteout" *Bull. Amer. met. Soc.* Vol. 27, p. 130, 1946

[2] COURT, A. "On the Arctic Whiteout" *Bull. Amer. met. Soc.* Vol. 27, p. 359, 1946

[3] STEFANSSON, V. *The Friendly Arctic;* The Macmillan Co., New York, 1921

[4] STEFANSSON, V. *Arctic Manual;* The Macmillan Co., New York, 1945

[5] MIDDLETON, W. E. K. "Note on the Visual Range of White and Grey Objects" *Quart. J. R. met. Soc.* Vol. 73, p. 456, 1947

[6] FRITZ, S. "Illuminance and Luminance under Overcast Skies" *J. opt. Soc. Amer.* Vol. 45, p. 820, 1955

[7] MIDDLETON, W. E. K., and MUNGALL, A. G. "Luminous Directional Reflectance of Snow" *J. opt. Soc. Amer.* Vol. 42, p. 572, 1952

(Note added in proof: Some data have recently been published by G. H. LILJEQUIST in *Norw.-Brit.-Swed. Antarctic Expedition, 1949-52, Scient. Results* Vol. II, Part 1A, p. 73, Oslo, 1956.)

ON THE RELATION BETWEEN METEOROLOGICAL CONDITIONS AND TOTAL AMOUNT OF OZONE OVER TROMSÖ

HARALD JOHANSEN
Vervarslinga på Vestlandet, Bergen

The variation in the total amount of ozone over Tromsö and associated changes in meteorological elements, such as height of tropopause, upper-air temperature and pressure, have been studied. The correlation coefficients between the total amount of ozone and the elements just mentioned show an annual variation with only small coefficients during midwinter, and highest correlation coefficients during April–May.

Correlation coefficients of total ozone with temperature and pressure at the levels of 3, 6, 9, 12 and 15 km show that the highest coefficients are found at 6 km.

Correlation coefficients between ozone and potential temperature in the stratosphere during summer show that the highest correlations are found at 12–14 km, decreasing upwards to the level of 18 km.

A comparison between the annual variations of the correlation coefficients and the annual variations of the horizontal ozone gradient between 60° and 70°N shows that it is impossible to attribute the ozone variations solely to horizontal advection.

By grouping the observations according to rise and fall regions of atmospheric pressure at ground level, it is shown that the variations of total ozone are related to the fall and rise regions at ground level, to the height of the tropopause and to the upper pressure variations, pointing to dynamical processes being of great importance to the connection between ozone and weather.

The variation of 5-day means of ozone in relation to the 5-day means of pressure and temperature at 6 km has been studied. It was found that the ozone content varies inversely to the upper cold troughs and warm ridges.

THIS investigation on the relation between meteorological conditions and total amount of ozone has been worked out by means of upper-air observations at Tromsö during the period May 1941–October 1944, together with ozone observations which have been carried out at Nordlysobservatoriet (Aurora-Observatory), Tromsö.

First, the correlation coefficients between total ozone, height of tropopause, higher tropospheric and lower stratospheric temperature have been found. As all the elements have pronounced yearly variations, the correlations have been computed by means of the deviations of daily values from the annual curves. The values thus obtained have been divided into six groups, each group containing the values for a period of two months.

Table 1 shows the correlations between ozone and tropopause height. The great difference in number of observations between winter and summer is partly due to lack of ozone observations during the darkest season, and partly due to the soundings reaching higher levels during summer than during winter. The correlation coefficients in *Table 1* do not indicate a very

Table 1. Correlation between Total Amount of Ozone and Height of Tropopause

Months	*n*	$r(O_3, h)$	*F*
1. December–January	86	−0·123	0·07
2. February–March	102	−0·434	0·05
3. April–May	181	−0·642	0·03
4. June–July	220	−0·524	0·03
5. August–September	190	−0·557	0·03
6. October–November	162	−0·317	0·05

strong correlation, but they show that some connection doubtless exists. The highest coefficients are found during the period April–September, with a maximum of −0·64 in April–May. The small correlation during mid-winter does not offer any reason for assuming a connection to exist between ozone and tropopause height. On the other hand, the fact that the correlation coefficient is negative might indicate a tendency in the direction just mentioned.

It has been shown by Tønsberg and Langlo that the maximum error in the ozone observations at Tromsö during the darkest period is about 12 times the maximum error during the period April–September. Therefore, great care should be shown when using the ozone measurements for the winter months.

In order to investigate the connection between ozone and temperature variations, the variation in ozone has been correlated to the temperature variations at 6 and 12 km. The height of the tropopause above Tromsö is on the average 9·4 km, with a mean annual range of 1·5 km. The lowest position of the tropopause which was observed was 6·4 km. The temperature variations at the level of 6 km can, therefore, safely be regarded as a good representation of the upper tropospheric temperature variation.

For the representation of stratospheric temperature variations, the level of 12 km has been chosen. This level is situated well above the mean tropopause height, but it is not sufficiently high to give merely stratospheric temperature variations. Owing to the rapid decrease in number of observations with increasing height, the level of 12 km was preferred and, consequently, the observations had to be rejected whenever the tropopause was above 12 km.

Tables 2 and *3* confirm the results given in *Table 1.* Thus, the correlations of ozone with the tropopause height, as well as with lower stratospheric and upper tropospheric temperatures, have an annual variation, showing moderate correlations during the period April–September with maximum

Table 2. Correlation between Total Amount of Ozone and Temperature at 6 gkm

Months	*n*	$r(O_3, t_6)$	*F*
1. December–January	122	−0·314	0·06
2. February–March	148	−0·481	0·05
3. April–May	206	−0·604	0·03
4. June–July	235	−0·601	0·03
5. August–September	223	−0·558	0·04
6. October–November	201	−0·448	0·04

Table 3. Correlation between Total Amount of Ozone and Temperature at 12 gkm

Months	*n*	$r(\theta_3, t_{12})$	*F*
1. December–January	29	−0·290	0·12
2. February–March	39	0·470	0·08
3. April–May	132	0·594	0·04
4. June–July	209	0·536	0·03
5. August–September	171	0·517	0·04
6. October–November	92	0·339	0·06

correlation coefficients in April–May, whereas the correlations during the darkest period roughly could be classified as hardly significant.

Table 4 gives the correlations of ozone with temperature, potential temperature and pressure at 3, 6, 9, 12 and 15 km. Owing to the small number of upper-air observations reaching heights of 15 km or more during winter, only upper-air observations during the period April–September have been used. For the determination of the correlation coefficients, 412 simultaneous upper-air and ozone observations are available. As in *Tables 1*, *2* and *3*, the coefficients have been determined by means of the deviations from smooth curves of annual variations.

The correlation coefficients between ozone and temperature show a maximum at the level of 6 km. The value −0·61 is in good agreement with the values given in *Table 2*, which gives a mean of −0·59 during the same period (April–September). The level of 9 km is situated just below the mean annual height of the tropopause (9·6 km) and belongs to the transitional zone between troposphere and stratosphere where the correlation between ozone and temperature changes from negative to positive values. Therefore, only small correlations are to be expected at this level of the atmosphere. The correlation coefficient at 12 km is +0·55. *Table 3* gives also a mean correlation coefficient of +0·55 for the same period.

Table 4. Correlation Coefficients of Total of Ozone with Temperature (t) and Pressure (p) at 3, 6, 9, 12 and 15 km, and with Potential Temperature (θ) at 6, 9, 12 and 15 km above the Surface. Number of Observations: 412

Correlated element	*Standard deviations*	*Correlation coefficients*		*F*
		Tromsö	*Oxford*	
t_3	4·338°C	−0·556		0·023
t_6	4·781°C	−0·605		0·021
t_9	3·659°C	−0·316		0·030
t_{12}	4·460°C	+0·552	(+0·56)	0·023
t_{15}	2·817°C	+0·521	(+0·52)	0·023
p_3	7·120 mb	−0·532		0·024
p_6	6·824 mb	−0·674		0·018
p_9	6·426 mb	−0·605	(−0·46)	0·021
p_{12}	3·504 mb	−0·561	(−0·43)	0·023
p_{15}	2·472 mb	−0·387	(−0·36)	0·028
θ_6	5·225°C	−0·562		0·023
θ_9	3·635°C	−0·080		0·033
θ_{12}	6·552°C	+0·578	(+0·60)	0·022
θ_{15}	4·976°C	+0·560	(+0·63)	0·023

Between ozone and pressure a negative correlation can be found at all layers, with the maximum value −0·67 at 6 km.

From the correlation of ozone with potential temperature it can be seen that the variations with height are almost similar to the variations of the correlation between ozone and temperature. Using the measurements from Oxford it was found by Meetham that the correlation between ozone and potential temperature increased with height, indicating a quite close connection at 18 km ($r = +0{\cdot}70$). In order to find a confirmation of the assumed increase of the correlation with potential temperature in the stratosphere above Tromsö, the soundings reaching the level of 18 km have been examined. The results are given in *Table 5*, which shows that the highest correlations between ozone and potential temperatures are found at levels of 12–14 km. Above these levels a decrease can be found.

Table 5. Correlation Coefficients of Total Ozone with Potential Temperature at 12, 14, 16 and 18 km

Year	*n*	$r(O_3, \Theta_{12})$	*F*	$r(O_3, \Theta_{14})$	*F*	$r(O_3, \Theta_{16})$	*F*	$r(O_3, \Theta_{18})$	*F*
1941	29	0·582	0·083	0·482	0·096	0·343	0·111	0·223	0·119
1942	57	0·702	0·045	0·756	0·038	0·670	0·049	0·544	0·063
1943	42	0·366	0·090	0·380	0·089	0·354	0·091	0·337	0·092
1944	50	0·490	0·072	0·520	0·070	0·351	0·084		
Mean values:		0·535		0·535		0·430		0·368	

Most investigators have regarded the ozone variations as resulting from redistribution of ozone already present in the atmosphere. The numerous studies of the connection between the variations of total ozone and other meteorological elements have pointed to the horizontal advection as an important cause of the day-to-day variations, since it was assumed that polar air always had a high ozone content and tropical air a low one. It was,

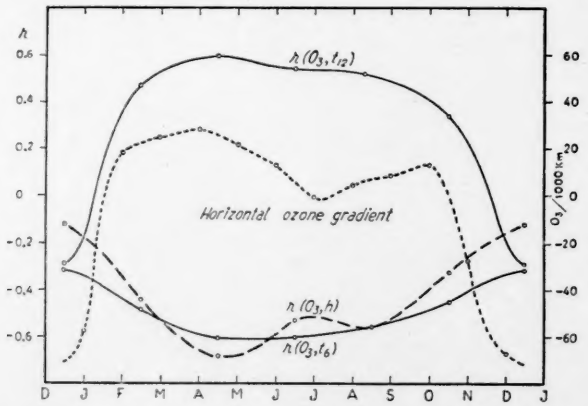


Fig. 1. Showing the annual variations of the correlation coefficients between total ozone and tropopause height (dashed lines) and between ozone and tropopause at 6 and 12 km (full lines). Dotted line shows the mean annual variation of the horizontal gradient between Dombås and Tromsö (units 10^{-3} cm O_3/1000 km)

however, early pointed out by Dobson and his collaborators that it was impossible to attribute the ozone variations solely to horizontal advection, and it was assumed that other factors played an important role.

Fig. 1 shows the annual variation of the correlation coefficients, together with the mean annual variation of the horizontal ozone gradient between Tromsö (70°N) and Dombås (62°N). From the figure it can be seen that the highest ozone gradient occurs during the period February–May with a maximum gradient in April. This corresponds well with the annual variation of the correlation coefficients having their maximum values in April–May. On the other hand, only small values of the horizontal ozone gradient are found during the period June–November, whereas the correlation coefficients show that, during the same period, some connection doubtless exists.

Several investigators have pointed to the effects of vertical motions. If the ozone content below the layer of maximum ozone content can be regarded as a semi-permanent gas, a uniform sinking of air from the layer of photochemical equilibrium would transport ozone from higher to lower levels in the atmosphere. At the same time photochemical processes in the upper layers would quickly restore the ozone deficiency, leading to an increase of total ozone. An ascending motion would give the opposite effect.

Another point which also has to be taken into account is the effect on total ozone produced by horizontal convergence and divergence which accompanies the vertical motions. This effect has been calculated by Nicolet, Dütch and Reed, who showed that it could be an appreciable one.

The connection between the large-scale vertical movements and ozone

Suppl.
Part.I
1956

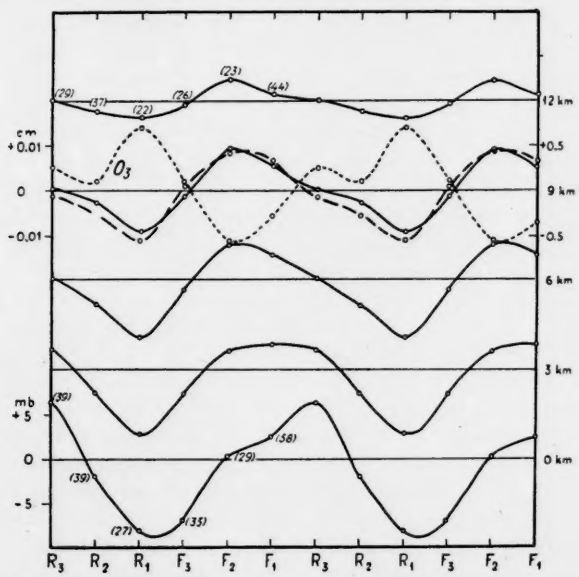


Fig. 2. Showing the anomalies of atmospheric pressure at the levels of 0, 3, 6, 9 and 12 km, together with the anomalies of the tropopause and the total amount of ozone. The groups F and R represent observations made in fall and rise regions of atmospheric pressure at ground level. Number in brackets gives number of observations in each group

deviations which has been found to exist shows that the dynamical processes must be of great importance, as Palmén and Haurwitz suggested about 20 years ago.

During a period of one year I have grouped the observations according to rise and fall regions at ground level, omitting the observations during the darkest period December–January. *Fig. 2* shows the anomalies of pressure at the levels of 0, 3, 6, 9 and 12 km (full lines), together with the anomalies of the tropopause (dashed line) and the total amount of ozone above Tromsö (dotted line). The groups F represent observations made in fall regions and R in rise regions of atmospheric pressure, in the way that the groups F_2 and R_2 represent observations in the central part of the falling and rising regions, respectively. The figure can be regarded as representing the mean conditions of low- and high-pressure systems moving from west toward east. The figure shows that the upper pressure system is displaced to the rear of the surface system, just as it must be expected in a system of west-east moving lows and highs. The relationship between the ozone deviations and the pressure waves shows that negative ozone deviations appear to the rear of the surface anticyclone and above the central part of the falling region. The positive ozone deviations appear to the rear of the surface cyclone above the areas of rising pressure. The irregular shape of the ozone deviation curve during the falling period would probably disappear if the number of observations were increased. The figure reveals in a striking manner the close relationship which exists between ozone deviations and upper pressure waves, and between ozone and variations of the tropopause. It shows that the ozone varies exactly inversely to the pressure at the upper tropospheric and lower stratospheric levels and to the variations of the tropopause.

In the ordinary eastward-moving cyclone the pressure changes at ground level can be regarded as a result of two strong effects below and above the level of non-divergence, almost cancelling each other. If the upper-air divergence is strong enough to overcompensate the convergence in the lower layers, there will be falling pressure at ground level and vice versa. The redistribution of air-masses in the lower layers of the atmosphere will have no effect upon the distribution of total ozone, whereas convergence and divergence in the upper troposphere and in the lower stratosphere will cause ozone variations as calculated by Reed. The resulting horizontal distribution of total ozone must, therefore, show high ozone values above the rising pressure areas and low values above falling pressure areas at ground level. In this way the resulting horizontal ozone distribution will be more or less symmetrical with the upper pressure waves. The variations presented in the figure are also in good agreement with the result in *Table 4*, showing the highest correlation coefficients between ozone and pressure at 6 and 9 km with a maximum of $-0{\cdot}67$ at 6 km. The figure thus indicates that the best correlation between total amount of ozone and pressure is found in the upper tropospheric levels.

Fig. 3 illustrates another example of the close relationship between ozone and the upper pressure waves. In this figure the variation of the five-day means of ozone is given together with the five-day means of pressure and temperature at 6 km. The observations during the darkest season are omitted. The five-day values have been computed by means of the deviations of daily

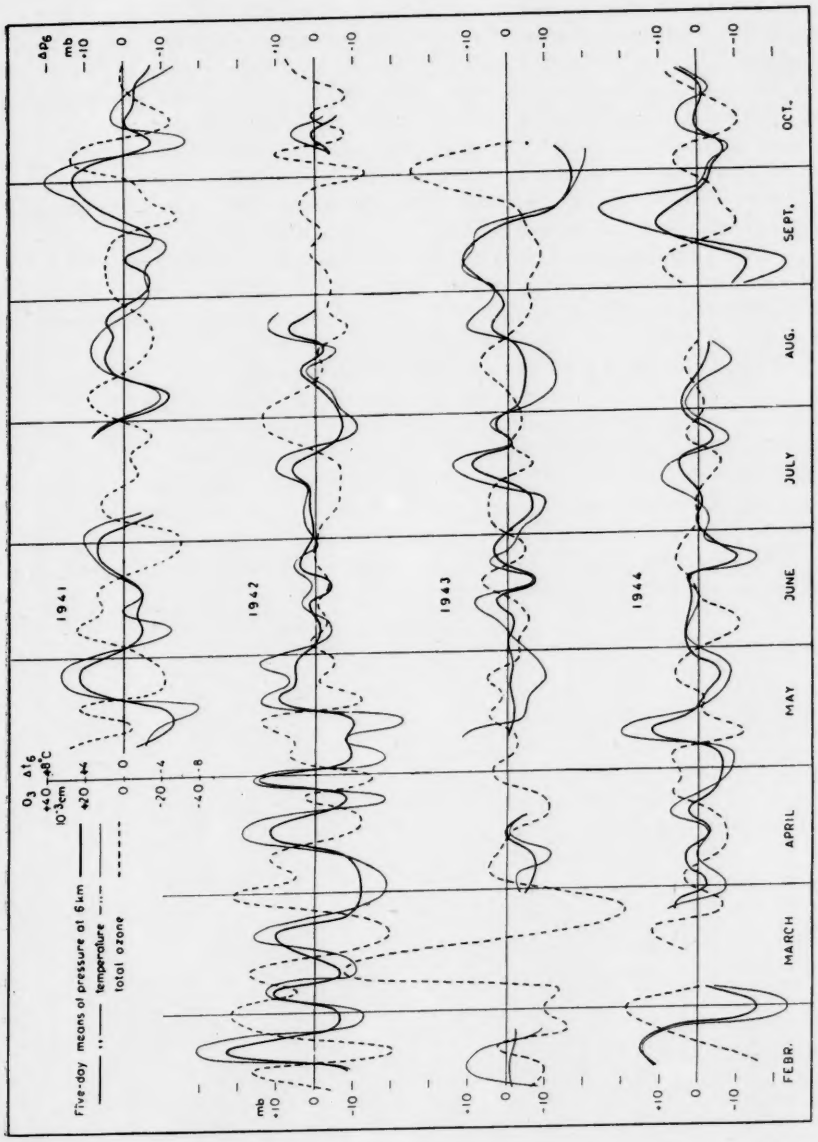


Fig. 3. Showing the variations of the 5-day means of total ozone together with the variations of the 5-day means of atmospheric pressure and temperature at 6 km

values from the annual curves. The full lines show pressure variations, thin lines are temperature and dashed lines ozone variations. Again, it can be seen that the ozone content varies inversely to the upper cold troughs and warm ridges. It also shows that the largest ozone variations occur during spring and autumn in connection with the strongest pressure systems, as would be expected if dynamical processes in the atmosphere play an important part in causing ozone variations. The figure also gives the impression that, by using 5-day means, a closer relation results than by using the day-to-day variations. In order to decide if this actually is so or not, the correlations of ozone with pressure and temperature was computed. The number of observations of not overlapping 5-day means are 166. The correlation coefficient between ozone and pressure is −0·65. The corresponding coefficient from the 412 day-to-day variations was −0·67. The coefficients between ozone and temperature was −0·61 and −0·63, respectively. Thus, the correlation coefficients computed by means of day-to-day variations and by means of 5-day values, are almost equal.

Finally, I should like to mention that Scherhag, in his book *Wetteranalyse und Wetterprognose*, has pointed to the similarity between the variations of total ozone and the variations of stratospheric compensation. Using the observations from Tromsö, the correlations between total ozone and stratospheric compensation were computed. Unfortunately, I have not the data here, but the computation gave about the same correlation coefficient as between ozone and temperature at 12 km. This result is quite reasonable since the variation of the compensation is proportional to the variation of the mean temperature between levels of 225 and 96 mb.

Supp
Part
1956

VISUAL RANGE IN THE POLAR REGIONS WITH PARTICULAR REFERENCE TO THE ALASKAN ARCTIC

J. MURRAY MITCHELL, Jr.
Office of Climatology, United States Weather Bureau, Washington 25, D.C.

The polar atmosphere often possesses a remarkably high optical transparency which can be attributed to very low dust and water-vapor contents. Below the tropopause, and especially in those regions which experience polar and Arctic frontal activity, however, extensive stratiform cloudiness usually develops. With a few notable exceptions, as when solid cloud masses have recently been sounded to heights of 39,000 ft (12,000 m) over the Arctic Ocean, cloud systems can usually be surmounted rather easily by propeller aircraft. Especially in the warmer months and near coastlines, mountains in the Arctic are submerged in heavy cloudiness.

Poleward of 70° and at altitudes from ground level to more than 30,000 ft (9000 m) pilots very commonly encounter "Arctic haze" in which visual range in the vertical is unimpeded but horizontal and slant visual ranges are restricted to an estimated 2–5 miles (3–8 km). Arctic haze is usually found in patches some 500–800 miles (800–1300 km) across and its presence can seldom be detected by ground observers. It is composed of very small particles ($\leq 2\mu$) which have not been certainly identified as ice, and appears to differ physically from the cirrus haze encountered by jet aircraft flying near tropopause levels in middle latitudes.

During the Arctic winter, deep inversions develop which, in the lowest few hundred feet above ground, are strongly super-refractive. Distant mountain ranges, which are normally on the horizon or even below it, sometimes "loom" into full view but terrain is seldom flat enough to allow the appearance of laterally extensive mirages.

At times of these deep Arctic inversions, and of the very low surface temperatures accompanying them, inhabited areas become shrouded in localized patches of ice fog. The copious combustion of fuel, together with the discharge of spent steam from central heating systems, has become so prevalent in Arctic settlements in recent years that the local atmosphere soon becomes supersaturated whenever temperatures dip below −20° to −30°F (−30° to −35°C). Average frequencies of ice fog at the larger Arctic air bases in interior Alaska have rapidly increased along with accelerated human activity, but in daylight hours are approaching the maximum frequencies which are likely to be permitted by climatological frequencies of favorable temperature.

Owing to very low specific humidities, radiation fogs are rather infrequent in polar regions. Advective fogs, however, are common, especially along coastal areas during the warmer season. Dense Arctic sea smoke may be favored along tidal flats during the colder season where pronounced tidal action prevents the water from freezing over solidly.

By way of improving the operational dependability of Arctic air bases, it is felt that local inhibition or dispersion of ice fog and Arctic sea smoke in certain areas of Alaska is a practical possibility and worthy of attention. Ice fogs can be mitigated by the suppression of local production of moisture, and/or can effectively be drifted away from runways by special utilization of horizontal-axis wind machines. Frequently occurring, supercooled fogs elsewhere in Alaska are amenable to dispersal by dry-ice seeding.

Introduction

The optical transparency of the polar atmosphere is often so high that pilots and explorers in the Arctic have occasionally been moved to remark on it in their logs. One's ability sometimes to see so far and so clearly can cause mountain ranges to seem deceptively close, and can provide a pilot with vistas of mountains and snow-covered or river-strewn wasteland which are likely to be more expansive than anything in his experience. These circumstances are especially prevalent in the colder part of the year, and can be attributed to refraction and to the very small dust and water-vapor contents of the polar air. The polar regions, however, are not without clouds, fog and other obstructions to vision which occur from time to time. In some respects, the origin and characteristics of Arctic hydrometeors are so different from those of their middle-latitude counterparts that they warrant special consideration. Accordingly, it is the chief purpose of this paper to discuss the principal factors which determine visual range in the polar atmosphere, and to describe in some detail those factors, such as Arctic haze and ice fog, which are peculiar to the high-latitude regions of the earth.

Visual Range Aloft

Arctic cloud systems

According to various determinations of the latitudinal distribution of mean cloudiness over the earth, as by Brooks,[6] the average value of 60–70 per cent found for the polar regions substantially exceeds that of the lower latitudes. This cloudiness is concentrated seasonally in the summer and autumn months, meteorologically in the regions usually affected by polar and Arctic frontal activity, and geographically over the oceans and along coastlines. Especially during the warmer months, mountain ranges in the Arctic are frequently shrouded in heavy cloudiness. Except at rather uncommon times during the summer when insolation is strong enough to allow general cumulus or cumulonimbus development, polar cloudiness is very stratiform. Owing principally to the low tropopause found in high latitudes, clouds seldom extend above 25,000 ft (7500 m) in the colder season or above some 35,000 ft (10,500 m) in the warmer season. Even in the disturbances along the polar and Arctic fronts, *solid* cloud layers seldom extend above 15,000–20,000 ft (4500–6000 m), and propeller aircraft can usually break out on top of these at the more manageable altitude of 12,000 ft (3500 m). In contrast to these average statistics, it is instructive to call attention to the weather situation which prevailed over Alaska and adjacent portions of the Arctic Ocean on 27–29 December 1955, when an unprecedentedly strong meridional flow set in aloft and drove very warm and moist air northward from the Pacific Ocean. According to a U.S.A.F. weather officer stationed in Alaska at the time:[7]

> ". . . The warm moist air penetrating the Arctic Basin resulted in clouds to great heights. The 'Ptarmigan' (Air Weather Service reconnaissance) Flight, depending on celestial navigation at those latitudes, was forced to abort when they could not get on top at 29,000 ft. A jet was sent up along the Arctic coast to find out how high the clouds went. He was still in them at 39,000 ft (12000 m), at which point he turned back."

Arctic haze

To get back to the average state of affairs, the "Ptarmigan" weather reconnaissance missions over the Arctic Ocean have been instrumental in providing us with a climatology of our northernmost latitudes. Perhaps the most surprising outcome of the "Ptarmigan" observations has been the high incidence of haze at flight altitudes which restricts horizontal and slant visibility to an estimated range of 2–5 miles (3–8 km) without impairing vertical visibility to an obvious extent. The characteristics of Arctic haze, based on discussions between the writer and several weather officers* who have flown the "Ptarmigans", can be summarized as follows. The haze is found during a very high percentage of the time that the weather of the Arctic Ocean is otherwise clear, and, being encountered at all normal flight altitudes (700-500-300-mb levels), appears to have considerable vertical depth. Owing to the widely varying conditions of illumination, nothing can confidently be said about the seasonal variation in its frequency. The horizontal extent of the haze is usually limited to segments of the air route which are 500–800 miles (800–1300 km) long, but sometimes stretches over the entire 2000 miles of the reconnaissance track. A curious fact about the haze is that it is seldom found over land, as in the vicinity of Greenland.[12] A ground observer can rarely detect the presence of Arctic haze, just as thin cirrostratus layers encountered by jet aircraft in middle latitudes are often indistinguishable from clear sky to an earth-bound observer. But the similarity between Arctic haze and thin cirrostratus seems to end there: (1) Whereas the cirrostratus appears milky-white from the air, Arctic haze possesses a grey-blue hue in antisolar directions and a reddish-brown hue in the direction of the sun. (2) Whereas the cirrostratus is confined to a definite layer in the atmosphere, Arctic haze possesses very diffuse and indistinct upper and lower boundaries and occupies, in the vertical, a considerable fraction of the Arctic troposphere. And finally, (3), whereas the cirrostratus is obviously crystalline when viewed at close range from an aircraft ("You can reach out and touch it"), and produces sun pillars, parhelia, halos and similar optical phenomena in sunlight which are characteristic of ice crystals, the Arctic haze seems completely amorphous to the eye and is almost always without optical effects, at least those which are sufficiently bright to be seen through aircraft windows. The color effects in the haze suggest that, like the low-tropospheric haze of middle latitudes, the constituent particles are no larger than about 2μ. This size is so small that, should the haze be made up of ice, the crystals would be very rudimentary. Since ambient temperatures in the haze are normally in the range of -30 to $-35°$C, it is yet possible that supercooled water could be present although, to the writer's knowledge, no aircraft icing has ever been reported in Arctic haze.

The relatively low altitudes of the highest clouds occurring in the Arctic under normal conditions probably aggravates the difficulty of visual aircraft interception alluded to a year or so ago by MACDONALD.[10] This problem consists in a reduction of the visual acuity of jet pilots flying at high altitudes when no clouds are present at or above those altitudes, and comes about by

* Including the following United States Air Force personnel: Major T. A. Aldrich, Major H. G. Garber, Captain I. S. Israël, Lt. P. Graubard, Lt. F. J. Sorensen, Lt. E. Streed.

Suppl. Part. I 1956

the tendency for human eyes to focus a short distance ahead when the sky is perfectly monotonous, even though the pilot is attempting to focus at quasi-infinite range as when he scans for visual contact with a second aircraft. When and where it occurs, Arctic haze also makes interception and rendezvousing more difficult. At altitudes below about 30,000 ft (9000 m) it directly reduces slant and horizontal visual range; it also renders the sky greatly more homogeneous in the optical sense and thereby invites the loss of acuity cited by Macdonald. At altitudes above 30,000 ft, Arctic haze usually lies wholly below flight level, and its principal effect is to monotonize distant ground or cloud features.

Visual Range near the Ground

Hyper-extension of visual range

Principally during the Arctic winter, very deep and strong inversions commonly develop above the ground. The latter being covered with snow and ice, and surface winds usually being calm in the presence of the inversions, there is no opportunity for the lowest levels of the atmosphere to gain significant dust content. Further, the very low temperatures achieved under the inversion permit only a very small water-vapor content, and encourage a downward flux of moisture in the lowest layers of the atmosphere which usually culminates in sublimation on to the snow surface without the formation of ice crystals or fog in the air. Under these circumstances, unusually large visual ranges are favored in the lowest atmosphere. Additionally, the magnitude of the inversion is often sufficient in the lowest few hundred feet to refract horizontal light rays around the curvature of the earth. *Fig. 1* illustrates a group of lower-tropospheric soundings for Fairbanks, Alaska, which attended surface temperatures of −40°C or less. The dashed line in the figure represents the critical magnitude of an inversion (near sea-level pressure and at a temperature near −30°C) which would allow a horizontal ray to be refracted with a curvature equal to that of the earth, and is based on the formula given by Neuberger.[11] Almost without exception among the selected soundings, the lowest few hundred feet are super-refractive to the same, or a greater, extent. Notably in the case of the Alaskan interior during the winter season, vast areas of flat, snow-covered tundra usually share the strong inversions observed at Fairbanks, and a ground observer is rather frequently favored to see distant mountains "loom" into view which otherwise are on, or even below, the horizon. This is very annoying in geodetic surveying. It is not often, however, that Arctic terrain is sufficiently flat to allow the appearance of laterally extensive mirages. The phenomenon of super-refraction is credited sometimes with lengthening the periods of daylight and twilight by several minutes over astronomical expectancy.

Blowing snow

Together with fog and ice fog, which are to be discussed presently, blowing snow constitutes one of the principal obstructions to visibility near the ground in polar regions. Being typically dry and composed of fine particles, snow in the Arctic is easily swept into the air by gentle or moderate winds. In the range of about 9–14 m.p.h. (4–6 m/sec), wind commonly

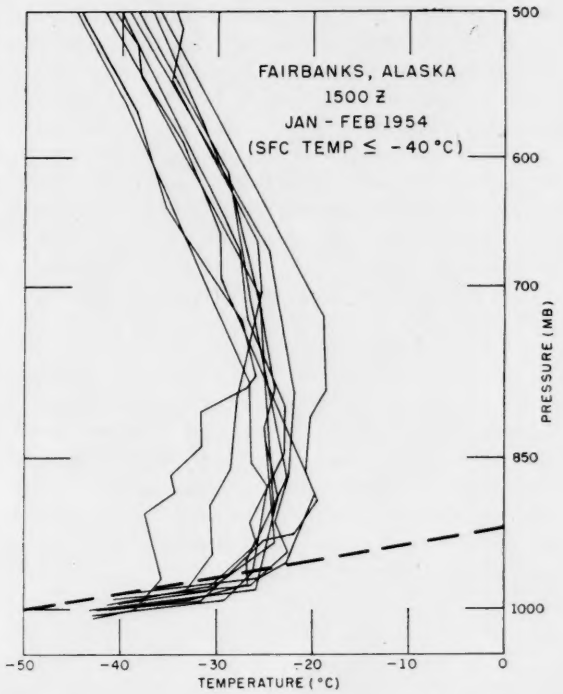


Fig. 1. Typical tropospheric temperature soundings at Fairbanks, Alaska, which accompany surface temperatures ≤ −40°C during the winter season. Dashed line shows critical magnitude of inversions near the surface which refract horizontal light rays around the earth's curvature

raises the snow crystals a few feet off the ground, and obscures many kinds of surface objects from a ground observer or from low-flying aircraft. With speeds exceeding about 15 m.p.h. (7 m/sec), wind can carry the snow to much greater heights and may obscure buildings, radio masts and entire airfields. These critical wind speeds are often exceeded at polar latitudes, and blowing snow is reported on an average of 5–10 per cent of observations there in the winter season. An example of excessive frequencies of blowing snow is that of Barter Island Air Force Base, Alaska (latitude 70°N), where average occurrence reaches about 25 per cent of all observations in November.

Coastal fogs

Owing to very low specific humidities, radiation fogs are rather uncommon in the Arctic. Especially during the warmer season, however, advective fogs often develop along extended segments of the Arctic coastline as well as over the larger, seasonably cool bodies of water. Coastal stations in Alaska, for example, experience advective fog some 10–30 per cent of the time during the summer. Islands in the Bering Sea, and some northernmost portions of the Canadian and Soviet Archipelago, evidently experience fog more than half the time in parts of the summer season.

During the winter, so much of the water surface of the Arctic is frozen

Suppl.
Part.I
1956

that the incidence of fog is much reduced from its warm-season levels. In the Sub-Arctic, however, Arctic sea-smoke (steam fog) develops when and where cold Arctic air-masses push out over the seasonably warm ocean. Normally, these outbreaks are accompanied by offshore winds which prevent the fog from drifting inland on a gross scale. Sometimes, during periods of extreme cold in the Sub-Arctic, local patches of Arctic sea-smoke develop over the inner reaches of fjords and tidal flats where the amplitude of sea-level fluctuation due to ocean tides is sufficiently large to prevent the formation of a solid ice sheet. Then, as fast as the fog can form (by evaporation from the warm water exposed to the air through holes in the tide-battered ice), it sometimes drifts over adjacent settlements. In the winter season, the Anchorage area of the Alaskan coast, for example, receives roughly a 10 per cent frequency of fog. Daily tidal variations of sea level in nearby Knik Arm average about 20 ft, and appear to be instrumental in keeping ice there partly broken up during all except the most severe winters. Aerial reconnaissance and observations of wind drift in the Anchorage area leave no doubt that the major portion of winter fogs there originate locally over Knik Arm and adjacent tidal areas. Being principally supercooled water fogs, they often transform the trees into beautiful fairyland forests by blanching them with rime. Unhappily, they also cause aircraft icing along with the usual complaint of low runway visibility.

Ice Fog

Thus far in the discussion various phenomena affecting visual range have been described which man has found *au naturel* in the polar regions. The phenomenon of ice fog, however, is one for which man himself is chiefly responsible. Because, in recent years, ice fog has evolved into a major operational headache at some of our large Arctic air bases, it seems appropriate to discuss the problem at some length.

Cause of ice fog

At the very low temperatures often occurring during the polar winter, the atmosphere is capable of retaining very little water vapor. Such low temperatures, moreover, are almost invariably accompanied by minuscule surface winds. Under these conditions it is clear that local sources of water vapor are conducive to the formation of fog. One source of water vapor at some of our air bases is illustrated in *Fig. 2*, which shows a portion of Eielson Air Force Base, Alaska, in relatively warm winter weather. Here, steam can be seen issuing from numerous points such as defective steam traps, air vents from humid rooms (kitchens, laundries, etc.), and auxiliary heating plants. The steam plumes shown in *Fig. 2*, which evaporate readily during sufficiently high temperatures, merge into a more or less persistent fog blanket at lower temperatures. Also, at sufficiently low temperatures, other sources of water vapor become important, as illustrated in *Fig. 3*. Taken of Fairbanks, Alaska, at a time when the temperature was low enough for the combustion of fuel to increase the humidity of the surrounding atmosphere, this photograph illustrates that the chimney effluent of homes and business buildings can contribute to fog formation.[13] When temperatures are sufficiently low, automobile and aircraft exhausts also produce fog: the theoretical

Suppl.
Part. I
1956

Fig. 2. Clear-weather view of some water-vapor sources at Eielson Air Force Base, Alaska[14]

(*Facing page* 200)

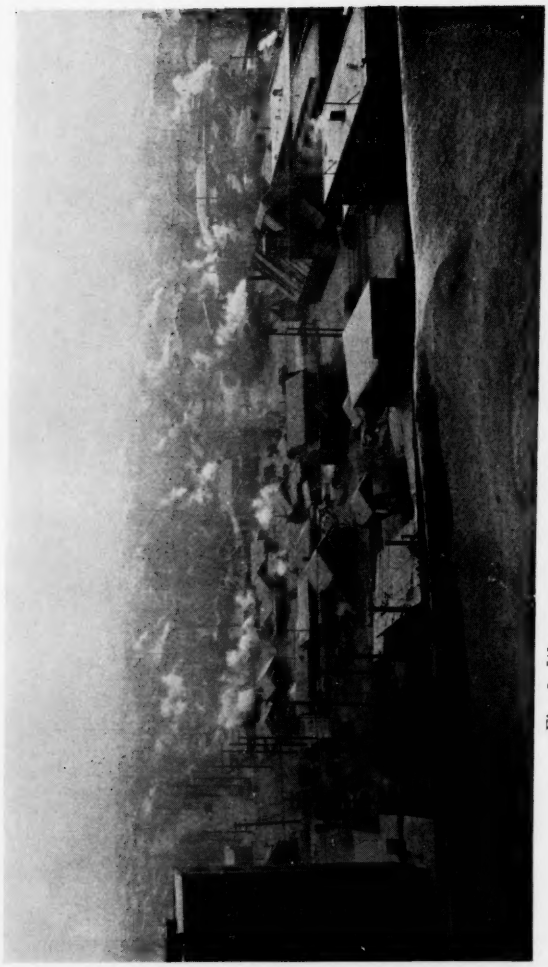


Fig. 3. Water-vapor sources and ice fog at Fairbanks, Alaska[11]

(*Facing page* 201)

Supp
Part
1956

relation between temperature and the tendency for hydrocarbon combustion to produce fog has been derived by APPLEMAN[2, 3] and is shown in his ice-fog forecasting diagram reproduced here as *Fig. 4*. In regions "I" and "II" of that diagram, conditions of temperature and humidity are favorable to the

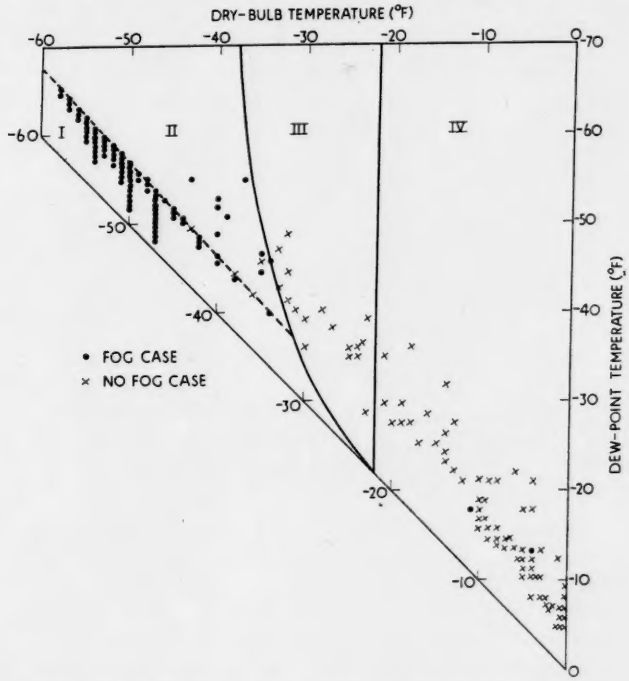


Fig. 4. Temperature-dew point diagram used in forecasting ice fog.[3] *Curve separating Regions I–II and III–IV gives critical combinations of temperature and humidity to the left of which hydrocarbon combustion supersaturates the atmosphere and to the right of which it decreases relative humidity. Observations for Fairbanks in the winter 1946–47 have been added*

condensation of effluent moisture even when all the heat of combustion is channeled into the atmosphere at the same place as the moisture.

Ice fog and temperature

In *Fig. 5* are compared three independent sets of data for interior Alaska which illustrate the observed relation between temperature and the overall tendency for ice-fog occurrence. Curve A, applying to the Fairbanks City Airport,[13] and curve B, applying to Eielson Air Force Base,[14] indicate the frequency of fog in which visibilities of half a mile or less are involved. The higher frequencies found for Eielson Air Force Base may be attributable to the relatively copious discharge of boiler steam into the atmosphere there at low temperatures. It may also reflect a universal increase in moisture production at human settlements (especially airfields) in recent years. (See ref. 14.) Curve C, which gives the total frequency of ice fog at Ladd Air Force Base irrespective of visibility in the fog, is obtained from heretofore

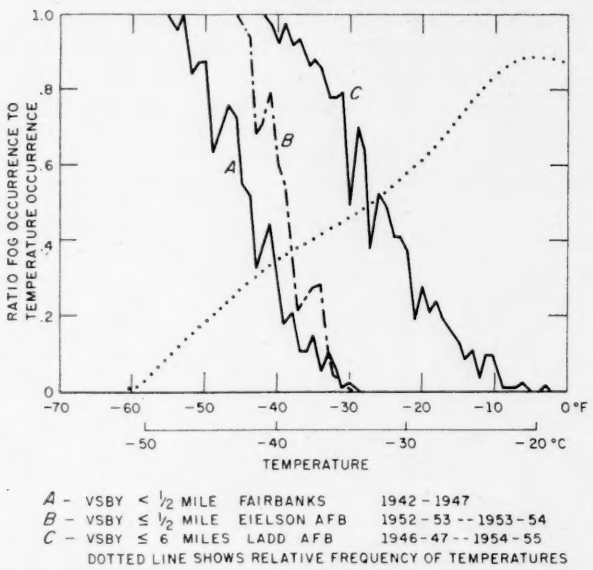


Fig. 5. Relative frequency of low-temperature (ice) fog as a function of temperature for three locations in the area of Fairbanks, Alaska

unpublished data of the United States Air Force Air Weather Service. The differences in slope between Curves A and B on the one hand and curve C on the other imply that the ice fogs which form below, say, −40°F are normally denser than those which form at higher temperatures.

The frequency of incidence of ice fog, then, is dependent upon the level of human activity as well as upon the meteorological frequency of critically

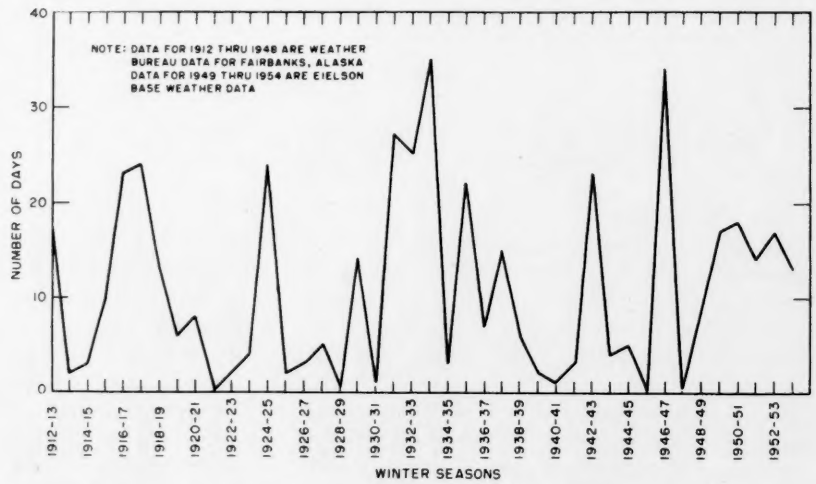


Fig. 6. Variation in number of winter-season days with temperature ≤ −40°C in the area of Fairbanks, Alaska[4]

low temperatures. *Fig. 6* serves to bring home the fact that the frequency of ice fog as determined by the meteorological situation must be a highly variable quantity from year to year.

Ice fog and increasing air-base activity

In terms of climatological averages, the probable effect on local ice-fog frequency of building or enlarging an air-base can be estimated in the manner illustrated by *Fig. 7*. Confining our attention to the left side of the figure, note the three lines curving sharply upward to the right: these represent the cumulative frequency of daily mean temperatures below any value, shown for three Alaskan stations in the month of January. Superposed on these is a family of vertical, dashed lines, labeled in per cent, which refer to critical ambient-air relative humidities derived by APPLEMAN[3] above which hydrocarbon combustion contributes to fog formation at the indicated abscissa temperature. A curve representing the cumulative frequency of ice fog at Ladd Air Force Base is added to the figure to illustrate its relationship to the curve of cumulative temperature frequency for the same station. Provided (*a*) there were no dispersion of relative humidity around its mean value of about 80 per cent (with respect to water) in the Arctic winter,[1] (*b*) all ice fog were formed by hydrocarbon combustion, and (*c*) winds were virtually calm each time the temperature dipped below −30°F, then the cumulative fog frequency curve should have the following shape: It should coincide with the cumulative temperature frequency curve to the left of the 80 per cent humidity line, and become a horizontal line to the right of the 80 per cent line. Owing to the failure of the above-mentioned conditions to apply universally, the curve of fog frequency breaks away from that of temperature to the right of the 0 per cent line, and then climbs with a gradually decreasing slope as an increasing fraction of ambient humidities fail to equal or exceed the corresponding critical humidity. The curve continues to climb on the right of the 100 per cent humidity line because, at those higher temperatures, fogs sometimes occur which originate from steam rather than combustion. A reasonable estimate of the incidence of ice fog which may result from the construction or enlargement of a base in the Arctic can be gained by use of the formula

$$\mathcal{F} = f_0\,(t \leqslant -39°\text{F}) + 0{\cdot}7f(-39° < t < -20°\text{F})$$

where $\mathcal{F}$ is the total percentual frequency of ice fog, and f_0 and f pertain to the relative frequencies of temperature t in the respective ranges indicated. The *maximum* frequency of ice fog to be expected at an inordinately large and active base would be, approximately,

$$\mathcal{F} = f'\,(t < -20°\text{F})$$

where f' is the percentage frequency of temperature below −20°F.

The right-hand side of *Fig. 7* shows the cumulative low-temperature frequency curve for Galena Air Force Base, Alaska, in each of the colder months of the year, by which one can rather easily estimate the probable incidence of ice fog were that air-base to be expanded beyond its present relatively inactive status. If we allow for the fact that such expansion could actually tend to inhibit fog formation by warming the air around the base

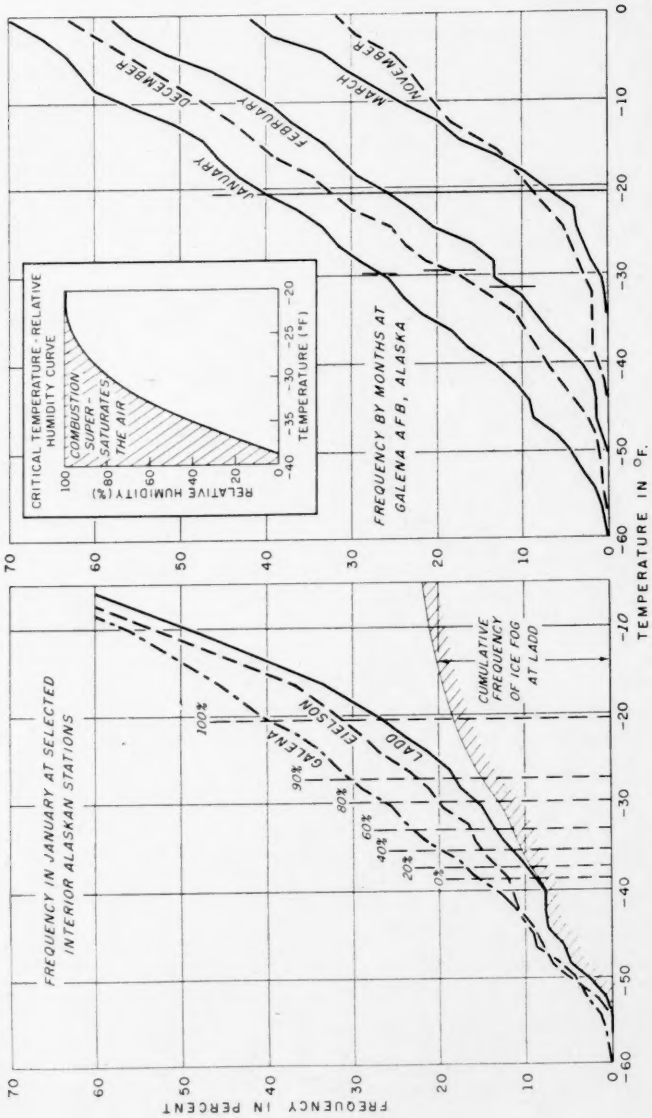


Fig. 7. Cumulative frequencies of low temperatures at three locations in central Alaska, derived from United States Air Weather Service "A and B" Climatological Summaries (approximate period of record 1942–55). Cumulative frequency of ice fog at Ladd Air Force Base in January has been added for comparison. For further explanation, see text

several degrees Fahrenheit, it can be concluded that the resultant frequencies of fog would rather closely parallel present fog frequencies at Eielson Air Force Base and Ladd Air Force Base.

Diurnal variation in ice-fog frequency

Very high solar zenith angles do not entirely suppress a diurnal temperature variation in the polar regions during the winter. In contrast to this temperature variation, the diurnal variation in frequency of ice fog is found to be very small in interior Alaska during January. If anything, ice fog tends to occur more frequently during the daylight hours at the larger bases, as shown in *Fig. 8.* This tendency can rather certainly be explained by the higher rate of activity on such bases during the day. Also shown in *Fig. 8* is

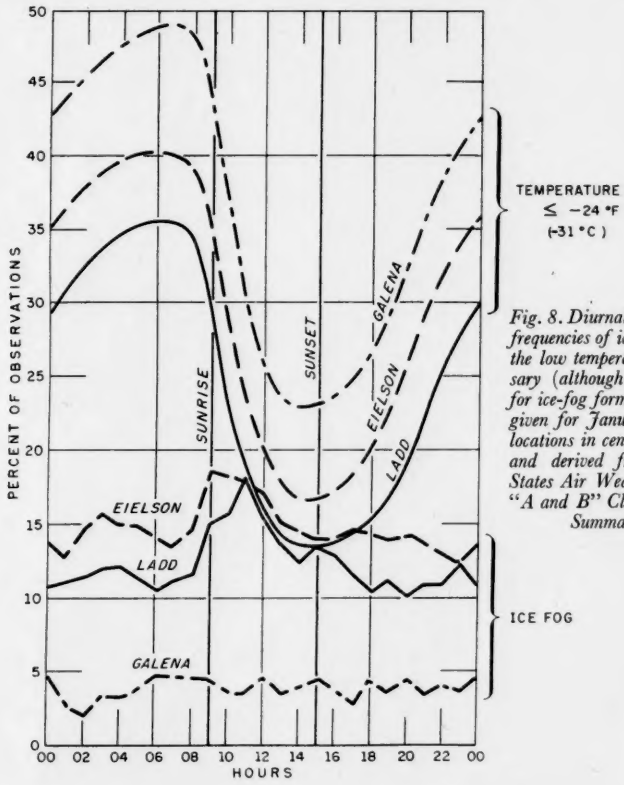


Fig. 8. Diurnal variation in frequencies of ice fog and of the low temperatures necessary (although insufficient) for ice-fog formation. Data given for January at three locations in central Alaska, and derived from United States Air Weather Service "A and B" Climatological Summaries

the estimated diurnal variation in frequency of temperatures less than or equal to −24°F for the same Alaskan bases. At any specified hour the temperature frequency curve may be considered as an approximate upper limit to the maximum likely frequency of ice fog under conditions of extreme activity and local moisture production. This statement is based on the approximation that

$$\mathcal{F} = f_0\,(t \leqslant -39°\mathrm{F}) + 0{\cdot}7 f(-39° < t < -20°\mathrm{F}) = f''(t \leqslant -24°\mathrm{F}),$$

Suppl.
Part.I
1956

where the *f*-symbols refer, as before, to the frequencies of temperature in the indicated ranges. According to the relationship between fog and temperature frequencies in *Fig. 8*, it is likely that the frequencies of ice fog at Ladd Air Force Base and Eielson Air Force Base have nearly reached their upper limits during the day-time, but could increase substantially at night if additional moisture were to be provided by accelerated base activities in those hours. Galena Air Force Base, on the other hand, has had a small fraction of the ice fog which could potentially develop there during any diurnal period.

Visual range in ice fog

The visual range in ice fog is characteristically very low, frequently narrowing to less than a quarter of a mile. Not only is the total particulate

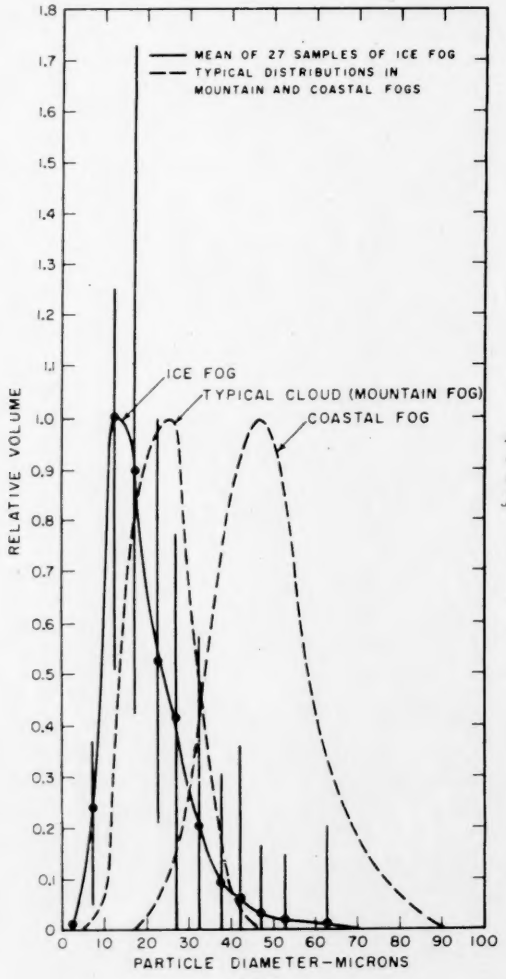


Fig. 9. Comparison of typical particle volume distributions for ice fogs and water fogs[14]

water content of ice fog comparable to that of high-temperature fogs,[14] but, as shown in *Fig. 9*, the average size of ice-fog particles is much less than of other types of fog and favors greater optical scattering. *Fig. 10* illustrates the cumulative frequency of low visibility in winter at the same Alaskan bases previously discussed. A very small fraction of the cases of low visibility in that region and that season is caused by factors other than fog. Especially in January, the low incidence of the lower visibilities at Galena Air Force Base, as compared with the incidence at Ladd Air Force Base and Eielson Air Force Base, is very apparent. The parallel slopes of the curves at the several stations emphasize the fact that, when ice fog occurs, it usually causes a complete deterioration from excellent visibility ($\geqslant$ 10 miles) to very poor visibility ($<$ 1 mile).

Fog in the Polar Regions: Remedial Measures

In its report on an investigation of ice fog, the Stanford Research Institute[14] has recommended a number of measures for alleviating the ice-fog problem at interior Alaskan air-bases. To generalize from these, the following points may be made:

(1) Steam-heating systems should be wholly enclosed, and provide for condensate return to the boilers.

(2) To the extent that (1) can be implemented, centralized heating systems should replace oil and coal space heaters wherever practical.

(3) Main heating plants should possess equipment (such as the glycol scrubbing system used at Eielson Air Force Base) which removes the moisture from the stack effluent.

(4) Scheduling of motor vehicle traffic should permit reductions in the number of vehicles operating during ice-fog weather and in the number of hours per day that vehicle engines must be left running.

(5) Motor vehicles which must operate in ice-fog weather should be equipped with water-vapor condensers such as those which have been designed for Arctic use by the Armour Research Foundation[9] to dehydrate automobile exhaust.

(6) Warm-up and take-off of successive aircraft should be separated as much as possible in both space and time.*

* The importance of aircraft operation to ice-fog formation can be verified by the following example. At a temperature near -40°C, at sea-level pressure, and with an initial relative humidity with respect to water of 80 per cent,[1] about 0·016 g/kg of water is required to saturate the atmosphere and another 0·056 g/kg in particulate form to reduce visual range to the neighbourhood of 1000 feet (300 m).[8] On the other hand, total normal fuel allowances for warm-up and take-off of aircraft in very low temperatures (as given in [14]) imply that local water production on a runway ranges upward from about 390 lb (175 kg) for the relatively small F-86F type jet aircraft during an average interval of 15 min prior to take-off. If this water is fairly realistically distributed through a volume of air corresponding to the lowest 120 ft over a runway one mile long and 600 ft wide ($\simeq 1\times10^7$ m^3), four F-86F aircraft taking-off about the same time would suffice temporarily to close the airfield at -40°C. Under similar assumptions, one B-50 aircraft can close the field, and a B-47 or a B-36 aircraft can temporarily maintain a dense runway fog despite the presence of ambient wind drifts of 1 or 2 ft/sec.

The Stanford Research Institute[14] has given a comparable analysis to demonstrate that the estimated total rate of water-vapor production on Eielson Air Force Base by automobiles, space heaters and steam heating units is quantitatively sufficient to account for the high incidence of ice fog there during the winter.

Suppl. Part. I 1956

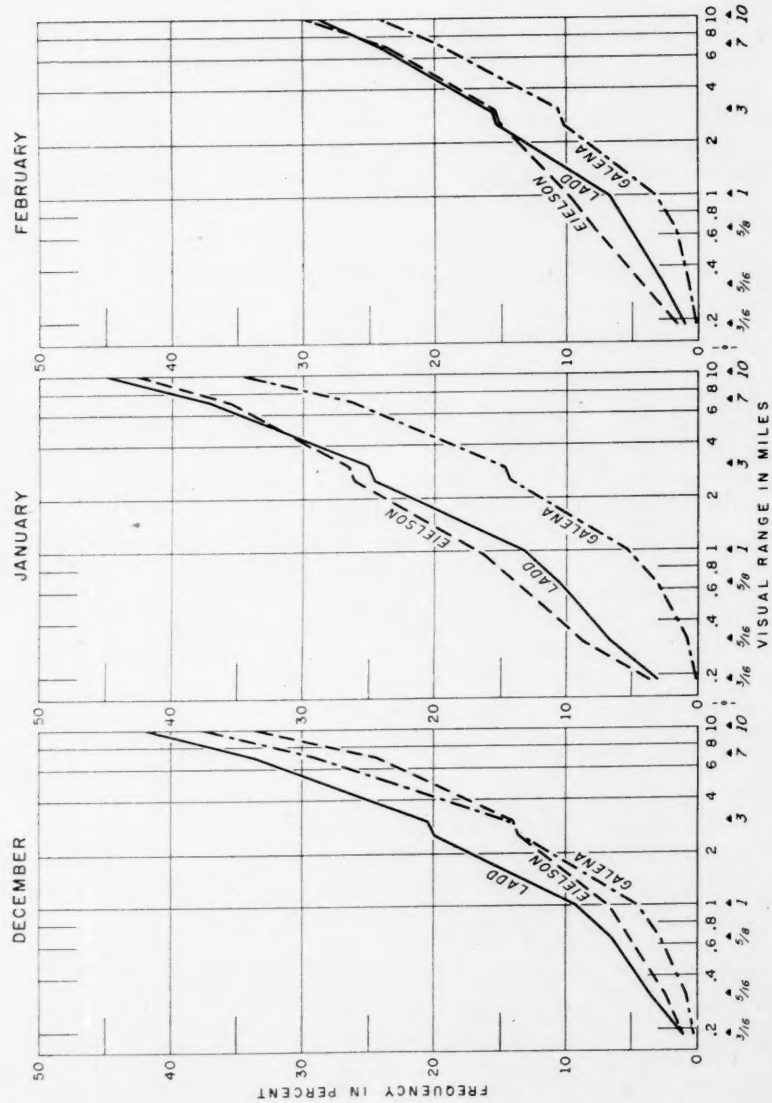


Fig. 10. Cumulative frequencies of low visibilities during three winter months at three locations in central Alaska. Data derived from United States Air Weather Service "A and B" Climatological Summaries

Suppl.
Part. I
1956

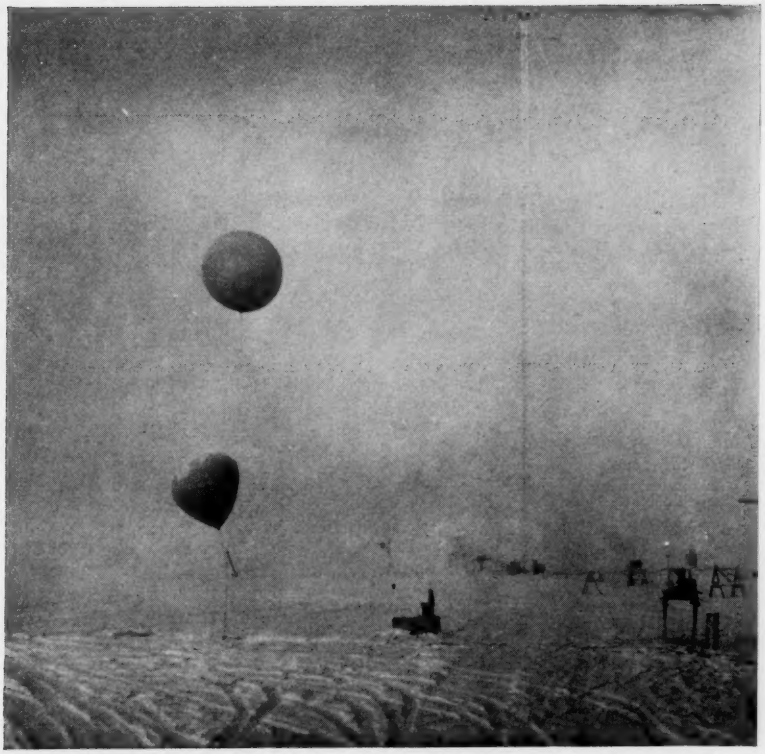


Fig. 11. Runway-level view within ice fog at Eielson Air Force Base, Alaska. Temperature about —42°C. Top of 30-m micrometeorological tower is in clear air above the fog layer.

Fig. 12. Ice fog over Eielson Air Force Base. Steam plume from power plant (right-center) often spreads out as a second discrete fog layer above the ground layer shown here[14]

(*Facing page* 208)

Supp
Part
1956

With regard to the planning of new installations, the following points can also be made:

(1) Selection of new air-base sites should be such that topographic depressions are avoided where Arctic air can stagnate, and such that drainage winds do not place them down-wind of other centers of population.

(2) Bases themselves should be laid out such that the runways are upwind of, and as far removed as practical from, the other base facilities. The runways should also be at a higher elevation than the other facilities wherever terrain permits.

(3) The buildings and other facilities of the base should be designed to emit a minimum of water vapor except at high elevations above runway level or downwind of the runway area.

To the extent that such measures are inadequate entirely to remove the threat of ice fog at air-bases, it seems worth while to consider methods of dispersing the fog by artificial means after it has formed. A number of methods have been suggested to accomplish this, but the physical or economic practicability of many of them has been seriously challenged. Among them are seeding by dry ice or silver iodide, which is sure to be ineffective against a fog known[14] to be initially composed of ice; and the use of FIDO, which releases so much water vapor into the atmosphere by combustion that it is sure to aggravate the fog instead of alleviating it under typical conditions of temperature. The mechanical mixing of the fog with overlying, warmer air has also been considered, but the economic factor seems prohibitive. Mechanical movement of the fog-laden air in horizontal directions, however, does not appear to be so demanding of power because the deep inversions which normally overlie ice fog suppress mixing of the fog layer with its superposing air. Fans along the runways of an air-base which are designed to induce horizontal motion of the fog-laden air would have merely to accelerate a volumetrically small air-mass to a speed comparable to that of the natural wind drift (less than 4·5 m.p.h. (2 m/sec) at Eielson Air Force Base), without expending the considerable additional power required to reshuffle a highly stable atmosphere. *Figs. 11* and *12* illustrate the limited depth and areal extent of an average ice fog in the Eielson area which make this "ventilation" system potentially practical. *Fig. 13* presents a number of soundings of the lowest layers of the atmosphere over Eielson Air Force Base during ice fog, and illustrates the extreme magnitude of the Arctic inversion at altitudes of 50–100 ft (15–30 m) overlying the fog layer, together with the less stable temperature distribution within the fog layer itself which should allow a reasonably efficient transfer of momentum from the fan drafts to the fog-laden air-mass as a whole. In an unpublished analysis of the problem in the Alaskan area, the writer has estimated that 15 400-h.p. machines per mile of runway, each with a blade wheel 15 ft (4·6 m) in diameter, would suffice to draw clear air over the runways of Eielson and Ladd Air Force Bases under more than 90 per cent of ambient wind conditions accompanying ice fog there.

With regard to the possible dispersal of winter-season fogs and Arctic sea-smoke which develop in certain coastal areas of the Arctic and Sub-Arctic, it has been observed that the vast majority of such fogs are supercooled droplet fogs which exist in ranges of temperature and wind speed

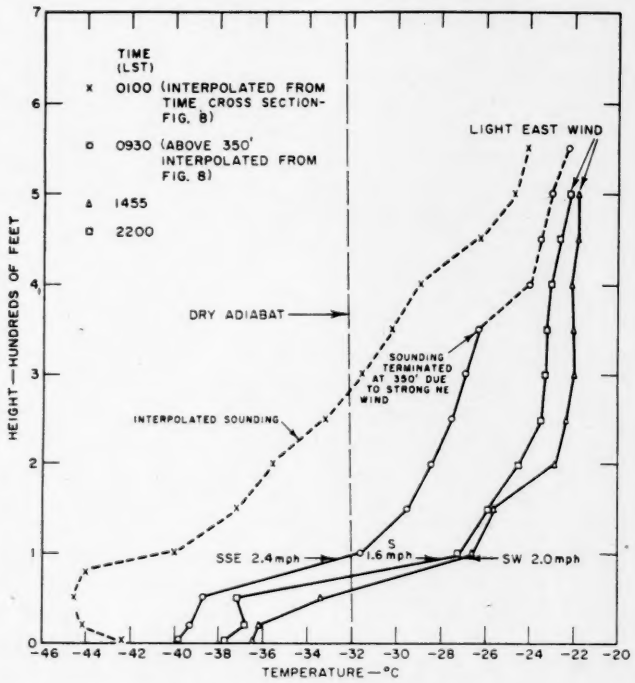


Fig. 13. Typical soundngs of temperature in the lowest 600 ft (200 m) during ice fog at Eielson Air Force Base, Alaska. Ice-fog layer was probably 50–80 ft deep against the ground. Data gathered with wiresondes.[4] *References to Fig. 8 apply to Bell's report*

almost ideally suited to seeding with dry ice or silver iodide. Despite these facts, very little intelligent effort has apparently been expended in the Arctic toward artificial dispersal of fog. And in view of the severe problem which fog presents to aviation in Alaska and other regions of the Arctic, the writer feels that a program for fog-dispersal operations should be given a trial there.

Acknowledgment

Grateful appreciation goes to Dr. Elmer Robinson and others of the staff of the Stanford Research Institute for their helpful discussions of ice fog, and for providing the writer with *Figs. 2, 3, 11* and *12. Figs. 6, 9* and *13* also are reproduced from the reports of the Stanford Research Institute. The writer also wishes to thank Dr. H. E. Landsberg for his review of the manuscript.

References

[1] Air Weather Service. "Climatology of the Arctic Regions" *Special Study No. 48*, Part I (see *Table 21*, p. 40), 1946

[2] Appleman, H. "The Formation of Exhaust Condensation Trails by Jet Aircraft" *Bull. Amer. met. Soc.* Vol. 34, No. 1, pp. 14–20, 1953

[3] ——. "The Cause and Forecasting of Ice Fogs" *Bull. Amer. met. Soc.* Vol. 34, No. 9, pp. 397–400, 1953

[4] Bell, G. B., Jr. "Environmental Aspects of Ice Fog Based on Field Studies at Eielson Air Force Base, Alaska" *Stanford Research Institute Scientific Report* No. VII; Stanford, Cal., 1955

[5] ——, and Robinson, E. "Wiresonde Observations during the winter of 1953–54 at Eielson Air Force Base, Alaska, *Stanford Research Institute Scientific Report* No. VI, AFCRC/GRD Contract No. AF 19(122)-634; Stanford, Cal., 1954

[6] Brooks, C. E. P. "The Mean Cloudiness over the Earth" *Mem. R. met. Soc.* (*London*). Vol. I, No. 10, pp. 127–138, 1927

[7] Ehrlich, Major A. U.S.A.F. Personal communication, 1956

[8] Houghton, H. G. and Radford, W. H. "On the Measurement of Drop Size and Liquid Water Content in Fogs and Clouds" *Pap. Phys. Ocean. Meteor. Mass. Inst. Tech. Woods Hole Ocean. Instn.* Vol. 6, No. 4, 1938

[9] Illinois Institute of Technology. "Research on the Elimination or Suppression of Ice Fog" Armour Research Foundation, Division of Engineering Mechanics Research, *Final Report*, 1952

[10] Macdonald, T. C. Report at Aeromedical Association Meeting, London, 21–23 March (abstracted in *Science News Letter*, 2 April 1955)

[11] Neuberger, H. "General Meteorological Optics" *Compendium of Meteorology*, p. 65, Boston, 1951

[12] Nilsentuen, Capt. R., U.S.A.F. Personal Communication, 1956

[13] Oliver, V. J., and Oliver, M. B. "Ice Fogs in the Interior of Alaska" *Bull. Amer. met. Soc.* Vol. 30, No. 1, pp. 23–26, 1949

[14] Stanford Research Institute. "An Investigation of the Ice Fog Phenomena in the Alaskan Area" *Final Report*; Stanford, Cal., 1955

Suppl.
Part. I
1956

SOME PROBLEMS OF METEOROLOGICAL OBSERVING IN POLAR REGIONS

N. J. SCHUMACHER
Det Norske Meteorologiske Institutt, Oslo, Norway

THE experiences which form the background for the following considerations were gained at Maudheim in the Antarctic (71°S, 11°W), but it is hoped that these comments will have some validity even in the Arctic.

It is well known that the weather conditions usually found in polar regions include: strong gales with heavy drifting snow and extreme cold with rime and ice formations. The low temperatures are often connected with great temperature variations, due to wave formations in the inversion layer. There is, however, another characteristic, which possibly is not so commonly realized, and that is the high intensity of the short-wave radiation during the summer half-year. There are two main reasons why this radiation is stronger in polar regions than in the medium latitudes. First, because of clear air, the intensity of the insolation is stronger, and secondly, the snow surface reflects most of the incoming radiation. This means that an object above the surface, such as a meteorological screen, is exposed to a short-wave radiation that usually is considerably greater than that which would be received in mediuin latitudes. No doubt, one of the greatest problems of obtaining reliable meteorological observations from polar regions arises from the presence of this radiation.

Naturally, the main problem of weather observing is to obtain correct values from the instrumental observations. An essential part of the standard instruments is placed in the meteorological screen, and it is thus of prime importance to use types of screens that are suitable for the prevailing conditions. In polar regions, the screen should meet two main requirements: (1) it should protect the instruments against drifting snow, and (2) it should allow sufficient ventilation to prevent radiation error in the summer half-year. In order to give full protection against drifting snow, the screen needs to be completely tight, since snow will penetrate the smallest opening as soon as the wind force exceeds a certain limit. On the other hand, in order to reduce the radiation error to a minimum on a clear summer day, the screen should be as open as possible. It seems that it is impossible to make a permanent arrangement which will fulfil the two requirements mentioned above. In the following it will be further clarified that these problems are of great practical importance.

Several expeditions have reported that their screen instruments have become packed with snow during gales, and thus put out of action. Especially the hygrometer and the hygrograph are easily put out of use, if snow gets in

the hairs. At Maudheim both these instruments were useless whenever a gale had been blowing for a few hours.

Although the drifting snow may result in few or no readings for some periods, it only seldom produces errors which cannot be readily detected. On the other hand, the errors caused by strong radiation are easily overlooked, and they are extremely difficult to correct for. On calm summer days at Maudheim we regularly took temperature readings on a ventilated Assmann psychrometer simultaneously with the observations in the screen. These observations showed that the radiation error in the screen might amount to 3° or 4°C, and for the last half of December the mean difference between the 12 G.M.T. readings in the screen and simultaneous readings on a ventilated thermometer was 1°C. When we compare the maximum values of the three-hourly readings, we find that the highest value in the screen was +3·3°C, whereas the correct value, read simultaneously on a ventilated thermometer, was −0·3°C. The highest value ever read on this thermometer was +0·2°C; this reading was observed on two separate occasions. On one occasion a gale was blowing, so that the screen value was exactly the same as that on the ventilated Assmann psychrometer, but on the second occasion the screen thermometer read +1·7°C.

Suppl.
Part. I
1956

Because of the difficulties which arise in connection with the meteorological instruments in the screen, one should seriously consider the possibility of using instruments which do not need a shelter like a screen. Thus, temperature measurements and recordings may with advantage be taken with electrical instruments. Artificial ventilation of the measuring unit can then be arranged more easily, and the recorders can be kept in heated rooms and thus be under control in all weathers. However, for some instruments such as hygrometers and hygrographs, there will probably always be a need for a shelter, which should be constructed so as to reduce the troublesome effects as far as possible. Because of the strong radiation, the walls of the screen ought to be made of open wicker-work, and artificial ventilation may be needed on clear summer days. Unfortunately, this latter arrangement requires continuous electrical supply, which is not always available on polar stations. It is possible that the ventilation may be improved by using a chimney, as did the Rønne Expedition to Marguerite Bay.[2] For use in bad weather, special covers should be available to place outside the wicker-work walls.

The meteorological element which is the most difficult to measure in drifting snow is precipitation. Since it is impossible to separate drifting snow from falling snow, it is difficult to devise a means for reliable precipitation measurements under such conditions. Probably the best solution is to mount the gauge higher than one would under normal conditions, since the frequency of drifting snow as well as its density decreases fairly rapidly with height. It may even be of advantage to place the gauge on a high mast or tower.

The more delicate parts of all instruments which are to be mounted out-of-doors should be covered as tightly as possible for protection against drifting snow. This must be kept in mind when anemometers and recording wind-vanes are selected.

The cold during winter in polar regions makes the standard instruments for humidity measurements unsuitable for use. It is well known that with

temperatures below some −20°C, hair hygrometers react very slowly to changing humidities, and the wet-bulb depression is too small to allow accurate readings with ordinary thermometers. Thermocouples may possibly be used with some advantage, since even small temperature differences can be measured with this instrument. However, it will often be difficult to obtain correct readings even on this instrument, because of the temperature fluctuations which take place in the inversion layer in winter. It is to be hoped that some new developments, such as the absorption hygrometer constructed at the U.S. Weather Bureau, will prove useful in polar conditions.[1]

The frequent rime formations on the instruments during the cold polar winter may cause some difficulties, but at Maudheim, at least, these troubles were generally less than had been anticipated. Both the radiation instruments and the anemometers were often rime-covered, but since this formation always occurred in fine weather it was comparatively easy to remove the rime before it seriously affected the readings.

The extreme cold makes necessary the special selection of oil for the recording instruments. All clock work to be used out-of-doors should be checked beforehand in cold chambers. The temperature fluctuations mentioned previously will often create difficulties for special investigations such as radiation measurements. When using instruments which are influenced by the temperature of the surrounding air, one has to follow the fluctuations and take the readings when the variations are the smallest. The rapid temperature fluctuations may often hamper the check readings of the radiosonde before the release. It is therefore of advantage to take these readings under cover, even in calm weather.

Another difficulty in connection with aerological investigations in polar regions is that of obtaining readings from higher levels in the winter half-year. Cooling of the batteries may cause the radiosonde to cut out before the balloon bursts. Special preliminary tests regarding these matters should be carried out with the radiosonde equipment which is to be used. An even greater problem is to obtain higher altitudes with the balloons before they burst, since, in extreme cold, the balloons burst at much lower altitudes. At Maudheim the mean height of bursting dropped from 21·9 km in the warmest month to 14·5 km in the coldest month. Improvement in this matter is probably contingent on development of better balloon fabric for use at low temperatures.

References

[1] Foskett, L. W. *et al.* "Infrared Absorption Hygrometer" *Mon. Weath. Rev.*, September 1953

[2] Peterson, H.-C. Weather Observing Program Compiled by Rønne Antarctic Research Expedition. Office of Naval Research, Dep. of the Navy, Washington, D.C., 1948

Suppl.
Part. I
1956

LOWER TROPOSPHERIC INVERSIONS AT ICE ISLAND T-3

A. D. BELMONT*
McGill University, Montreal, Canada

The complete record of the first two years of radiosonde observations from the Polar Basin have been analyzed with respect to thermal inversions. An objective classification system was introduced to permit exact determinations of type, magnitudes of component layers, and frequencies of occurrence, of inversions. The method has general applicability to other portions of the radiosonde ascent. A new unit of inversion intensity was introduced and appears to be an improvement over the use of lapse rate. Some new relationships of inversion characteristics are presented, and a general qualitative theory is suggested to explain the differences found with previous theory.

I. INTRODUCTION

A. The problem

Surface meteorological data from the Polar Basin have always been scarce and this is even more true of upper-air data. In 1950, dropsondes from 500 mb were begun on aircraft weather reconnaissance ("Ptarmigan") flights from Fairbanks to the Pole, and in 1952 the first series of surface radiosonde ascents from the central Arctic was started. The entire 23 months of these radiosonde records from the ice island T-3 were available for analysis.

The inversion was selected for study because it is one of the most characteristic features of the structure of the Arctic atmosphere. In addition, this type of study might yield results of practical usefulness in routine forecasting, and further, may provide some basic data for future comparison with International Geophysical Year measurements.

Upon review of the large literature on inversions it was found that most work has been in the field of micro-meteorology, based on special detailed data, and often extended only to about 300 m. Many of those who studied higher levels had too few observations, or readings too far separated vertically, to obtain a careful picture of the entire inversion layer. Our data were insufficient for any kind of micro-study but seemed as well suited as could be hoped for (with modern observational techniques), for the examination of the lowest 3 or 4 km. Since the frequency of observations was generally twice a day, a detailed study of diurnal variations was not possible, but as there was a relatively long period of data the series lent itself best to a discussion of monthly or seasonal changes.

Although the general causes of inversions have long been recognized, the quantitative relationships of the contributing variables are still unknown. Previous results on the subject are difficult to compare since all writers have implied different definitions and have measured characteristics differently.

* Now with Atmospheres Section, General Mills, Minneapolis, Minn.

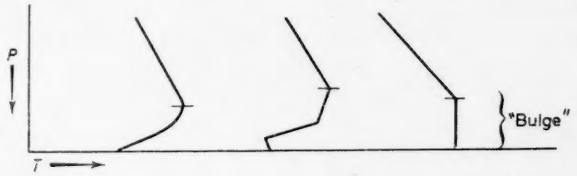
What now seem needed are: (1) a clear definition of the inversion layer, and (2) a method of classification into objectively determined types.

Only with these do comparable descriptions of inversion types with respect to frequency and magnitude of their components then become possible. Future comparisons of data may better explain what occurs in the atmosphere and help to enable the quantitative determination of each effect.

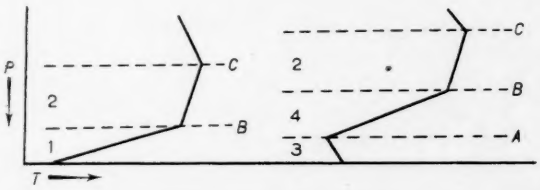
We offer here a possible definition and classification method and have applied them to a study of the T-3 inversions.

B. Basic definitions and abbreviations

Bulge—That lowest portion of the troposphere where the lapse rate first reaches or exceeds −3·5°C/km. Its characteristic appearance is indicated on a *P–T* diagram as below: this paper is a study of the temperature patterns in this "bulge" layer.



Inversion—In this study, broadly refers to the same layer as that defined above by the "bulge". In a restricted sense it is applied only to the B layer in which temperature increases with height with a minimum lapse rate of 3·5°C/km. In the literature, the component layers may be identified as follows:



1. Surface, ground or "nival" inversion
2. Main or upper inversion layer
3. Mixing or cold layer
4. Lower inversion layer

A, B, C. Layer boundaries defined in this paper (see II, C).

Lapse rate—For convenience of association with positive temperature changes, an increase in the temperature with height is considered to be a *positive* lapse rate, despite current usage in many, but not all, countries to denote it as negative.

Lower troposphere—From the surface to about 700 mb, or to the top of the "bulge", whichever occurs first.

"*Raob*"—Contraction for radiosonde observation. Also referred to simply as "ascent" or "sounding".

Subscripts	Refer to the layer or level indicated
Units	Metric units are used throughout unless specifically indicated otherwise
A,B,C, layers	A, B and C levels mark the top of A, B and C layers (see II, C)
H	Height
ΔP	Is always negative; minus signs are usually omitted
ΔP_a	The pressure change in the layer below the A level
P_c	Pressure at the C level
R_1	Lapse rate, $\Delta T/\Delta P$ or $\Delta T/\Delta H$, here considered positive when ΔT is positive
R_2	Inversion "intensity" $(\Delta T)^2/\Delta P$
RH_1	Surface relative humidity
T_1	Surface temperature
ΔT_b	By definition is always positive; hence signs are usually omitted
W_1	Surface wind

C. Résumé of literature on inversions

Although inversions were well known among residents of the Alps even before Hann first described them in 1870, or when Kerner related them to the local mountain winds in the Alps (1876), it was only in 1949 that a statistical treatment of alpine inversions, based on an ideal series of frequent radiosonde ascents, was published.[9] In the meanwhile, before radiosonde observations became available, several features concerning inversions elsewhere were recognized from kite and balloon ascents. One of the earliest balloon investigations of the upper levels in the Arctic was sponsored by the Prince of Monaco in 1906,[18] and it yielded possibly the first reference to the mixing layer frequently found over the sea.

In 1928 Evgenov[10,11] noted that the continuous winter inversion disappears in April along the Arctic coast, and in 1931 he noted from the 1914–15 kite ascents on the *Taimyr* that the increase of temperature in the inversion layer became greater as the surface temperature decreased.

By the early 1930s Moltschanoff instruments were in use along the Arctic coast. GUTERMAN[16] gave the results of 200 ascents during 1932–35 at Tikhaya Harbor (81°N, 60°E); TAUBER[38] discussed Dickson Island (74°N, 80°E) raobs for 1935–37; and KONDYRIN[23] treated Cape Schmidt data of 1935–36 whose inversions he classified into three types: surface, base above 200 m, base above 500 m. PENNDORF[26] summarized aircraft and kite ascents of Little America I and II for the summer months, and SCHNEIDER-CARIUS[31] reviewed Barkow's original 1911–12 kite data from the German Antarctic Expedition.

In 1925 and 1933, Sverdrup published the results of the *Maud* expedition's 1918–25 kite data along the north-east coast of Siberia. His most interesting deduction was that July would show a secondary maximum in both the thickness and the temperature difference with respect to the spring and fall months, although he had too little evidence to prove it from his three July ascents. His explanation was that the free air temperature continues to rise after the temperature at the melting ice has reached a constant value of 0°C. An inversion develops due either to warming of the air some distance above

the ice by direct absorption of heat, or to advection of warmer air from other regions. However, in spring and fall, at below freezing temperatures there are convection currents above the snow-surface when it is heated by solar and sky radiation. The surface temperatures are higher than air temperatures a short distance above the snow which results in mixing in the lowest layer. Sverdrup's results contrast with Evgenov's, which showed no mixing layer in winter, while the *Maud*'s ascents always showed one.

POAGE[29] carefully summarized the first two years of "Ptarmigan" dropsondes (1950–52), which represents the first upper-air data for the central Arctic between the surface and 500 mb. His data confirmed Sverdrup's expectations above. The dropsondes showed frequent disappearance of the inversion in the autumn and spring months, although the *Maud*'s soundings always recorded it or were terminated below what Sverdrup assumed to be the base of the inversion. Also, "Ptarmigan" observations showed much less sharp inversions than those of the *Maud*.

In a comparison of inversion data based only on mean constant pressure levels, BELMONT[3] showed that ice island T-3 had the most intense temperature increase (+13°C) (difference between surface and maximum mean temperature aloft), of all the stations used. These included Alert, Eureka and Little America III for the winter months from December to February. However, the latter station had a March mean inversion of 19·7°C, higher than any other station. T-3 most closely resembled Eureka with respect to inversions.

All the above references refer to maritime Arctic and coastal Arctic locations. Two important references on continental Arctic inversions are those by FLOHN[13, 14] who studied Siberian data, especially for Yakutsk, and BEDEL[1] who summarized the results of the French Greenland Expedition of 1949–51.

FLOHN[14] introduces the term "nival" inversion to distinguish the winter surface inversion due to radiation from a snow or ice cover, from the maritime type which contains a mixing layer due to wind and turbulence. He agrees there is little essential difference between this mixing layer type and the normal trade inversion.

Flohn disagrees with WEXLER[43, 44] who, on the assumption of clear and calm conditions, expected to find that as the surface temperature *decreased* below a critical temperature which was attained in a matter of hours, the surface inversion *decreased* and the upper isothermal layer became deeper. Flohn's Yakutsk data disproves this theory on both counts: the ground thermal inversion *increases* as surface temperature *decreases*, and the height of the isothermal layer is independent of surface temperature. Under IV our data will confirm Flohn's interpretation.

For Yakutsk, Flohn, who also uses constant pressure-level data, finds a 5-year mean January temperature increase of 16·6°C from the surface to 800 mb. Extreme mean values of 20 or 25°C near Verkoyansk may be expected due to the advection of warmer Pacific air aloft. He finds that both the height and the magnitude of the inversion increase from the north Siberian coast toward the interior. Also, it appears that inversions are not dependent upon pressure systems or air-masses, but rather upon radiation and history, and he concludes that the less the cloudiness, the greater the temperature increase in the inversion.

Supp
Part
1956

In the only other study made of continental Arctic data, Bedel presents, for the central station (Eismitte) in Greenland, monthly mean ascents which are based on constant pressure levels at 50-mb intervals, from which mean inversions of only +2 to +4°C/100 m can be derived. His example of a typical Arctic air-mass, however, shows an increase of +13·9°C/100 m. He concludes that wind speed > 10 m/sec^{-1} is a necessary but not a sufficient condition for the winter surface inversion to disappear. Flohn points out two extreme thermal gradients: at Little America III of +110°C/100 m for 15 m (or +68°C/100 m for 35 m), and at Eismitte of +15°C/100 m. From Bedel's data, it can be seen that on 22 February 1950 the latter station experienced an even stronger gradient of +17·7°C/100 m for 7 m.

Both Flohn and Bedel recognize that they cannot fix the top of the inversion exactly due to the infrequent vertical distribution of radiosonde values, and Flohn has assumed the top of the main layer to be where the lapse rate reaches −3°C/km.

Among the literature on inversion theory, one must mention again the work of Sverdrup and Wexler, Brocks[4] and the standard texts of Petterssen and Haurwitz. The remarkable two-hourly sequence of ascents for a 24-hr period of Ekhart[9] is an excellent example of alpine inversion evolution under ideal conditions.

On the classification of soundings, several recent papers deserve attention. The efforts of Schneider-Carius[31, 32] take the form of an association of vertical temperature and humidity profiles, with clouds and weather. He classifies the tropospheric temperature distribution into four general types, of which one is an inversion type. He considers the lower inversion layer as being a fundamental planetary layer of the troposphere. A new approach is made by Pône[30] who suggests an objective air-mass classification based not on its origin or temperature, but rather on the evolution in time of the entire sounding. He associates the types with synoptic patterns rather than with geographic sources.

One of the greatest obstacles toward advancement of upper-air knowledge has been the difficulty of processing radiosonde data. A recent and revolutionary idea to simplify this task is the method of graphical presentation of upper-air data in terms of unitary processing, devised by Bellamy.[2] So far, only qualitative representation has been achieved, but probably quantitative evaluation will soon become possible also, which should then permit more rapid processing of the data already existing.

We will only mention, for the sake of comparison, that despite the 100 years since the discovery of the subsidence-type trade inversion,[28] relatively little has been done toward its description. Von Ficker's classic paper,[12] describing the *Meteor*'s 1925–26 observations, does not seem to have been supplemented by new data except along the South African coast[37] and along the eastern Pacific.[41, 25] Certainly, the large-scale areal variations in the pattern of oceanic inversions, including the polar maritime type, should be investigated further.

D. Ice island T-3

Ice island T-3 is believed by many to be a portion of formerly land-fast shelf ice, probably from the northern coast of Ellesmere Island. This large

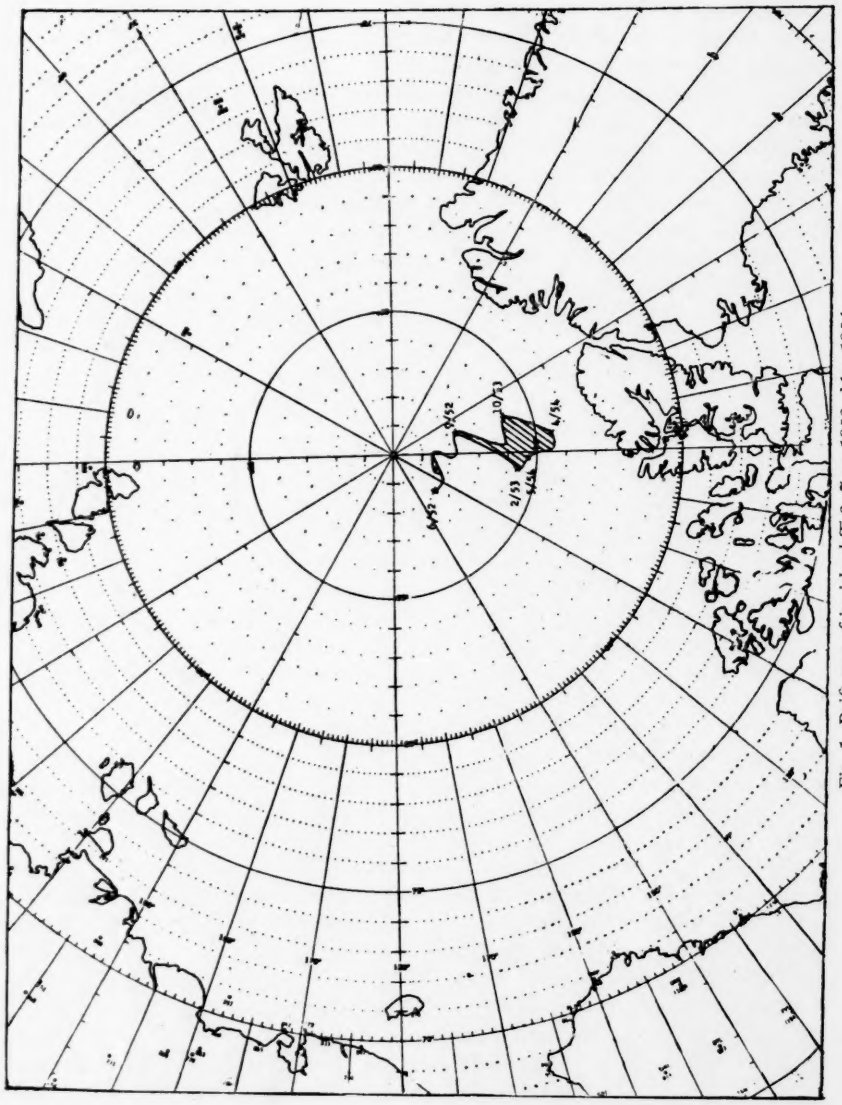


Fig. 1. Drift area of ice island T-3, June 1952–May 1954

"floe" is about 5 by 10 miles in area and some 200 ft thick, and has a maximum height above sea level of about 40 ft. The station height was 20 ft (6 m). Its surface undulates in a series of ridges and troughs varying from 100 to 1000 yd apart, which in summer become drainage canals for melt water.

The Russians first established a station on drifting ice in 1937–38, but this was on ordinary, relatively thin, pack-ice which broke up beneath them as it drifted into the Norwegian Sea. Following the discovery in 1947 of the first ice island, it first became possible to install permanent stations for scientific observations. These islands seem to drift in circular paths around the Arctic Basin. In April 1952, the U.S. Air Force established a geophysical station on T-3, the smallest of three major ice islands known at that time. It was evacuated in 1954 when the station drifted so close to the weather stations on Ellesmere Island that the information being gained was insufficient to compensate for the great difficulty and expense of maintaining the station. The Russians have continued to establish mobile research stations on the pack-ice in the central Arctic and to maintain each one for many months. So far as is known they have not occupied any of the large ice islands.

For the period of occupation, the drift area of T-3 is shown in *Fig. 1.* Within the hatched areas its course was erratic, and recrossed itself often. The total range in position of about 4° of latitude was not considered sufficient to warrant separating the data by different locations, so all data are summarized as if they referred to a single place.

For a discussion of the surface weather at the Russian stations, MEYER[24] has made a study based on radio and teletype reports as received in Germany. He estimated that only about 60 per cent of the total observations were received. As we did not receive surface records from T-3 in time to process them, the only information about surface weather given here is that contained in the frequency tables *6–11.* These reflect only the weather at the time of radiosonde ascents, taken at a maximum frequency of twice a day.

We refer the reader for other very interesting, but here unessential, details about ice islands to summaries by ZUBOV,[45] DIBNER[8] and KOENIG.[21]

II. DATA

A. The original material

1. *Data available*—About 1100 radiosonde observations for ice island T-3 from 21 June 1952 to 26 April 1954, in the form of microfilm of the original adiabatic charts (WBAN 31) used in computing the ascent data, were the raw material to be analyzed. The distribution of observations over the period are given in *Table 5,* and the general location of the station during the period is given in *Fig. 1.*

It will be noted that observational representation varied greatly over the 23 months. The poorest months, May and October, have only about 50 per cent of the best months. June has 70 per cent, and February, August, September and November each have about 85 per cent of the 5 months with excellent coverage. This distribution seriously weakens the equinoctial seasons in later comparisons of relative monthly frequencies of various inversion types and parameters, and should therefore be remembered. We prefer not

Suppl.
Part. I
1956

to adjust by a weighting factor any of the actual frequencies, for we ought then also to adjust for many other relative frequencies of variables, so that final interpretation would be more hindered than assisted. We prefer to use actual frequencies throughout, and to apply mentally any qualitative adjustments desirable as each individual comparison may require. The assumptions involved in such adjusting frequently introduce as much error as is implied by the original distribution. It is easier to interpret correctly the biased, but actual, frequency data than to evaluate the hidden factors in "corrected" frequency data.

Observations were attempted at or near the standard hours of 0300Z and 1500Z daily, but there were frequent omissions, as usual, due to extremely severe operating conditions for both equipment and personnel. Observations in strong winds were seldom possible, although they were made successfully in 16 m/sec winds (35 m.p.h.). In general, the data are excellent and the U.S. Air Force personnel who observed them deserve high praise.

All ascents in the original collection which were marked "doubtful" by the observers have been eliminated, as were those which were partly incomplete at any level below the top of the inversion.

The data sheets provide pressure-temperature values for all necessary levels, and heights only for standard isobaric surfaces. For practical reasons, it was decided not to interpolate for heights of significant levels but to use P-T relationships throughout. Due to space limitations on the IBM card, upper-level winds and dew points were not included in the analysis material. The ascents include observations of surface temperature, pressure, relative humidity, dew point and wind at the time of ascent, but contain no information on cloudiness or weather. Only well after all data had been processed on to IBM cards was microfilm of the surface weather observation forms received. As it was then impractical to process the surface material, only daily résumés of the cloudiness, weather and strong winds were prepared.

The only available series of synoptic maps for the Arctic area was the once-daily surface and 500-mb German *Täglicher Wetterbericht*.[15] The frequent and rapid change of pattern in the Arctic severely limits the usefulness of once-daily charts; indeed they may lead to seriously erroneous impressions of the pressure pattern development.

2. *Data accuracy*—The American radiosonde unit model AN/AMT/2(b,c) together with direction-finder SCR-658 were used during the entire period. The temperature element (ML-419/AMT-4) was of the exposed white thermistor type without any shielding. Any instrumental radiation error should be relatively insignificant except during the transition periods when ascents are taken alternately in direct sunshine and darkness. The humidity element (ML-418/AMT-4) is a lithium chloride strip. The instrument is described completely in ref. 40.

Because the modern radiosonde record is discontinuous, it is really not well suited for detailed soundings of the atmosphere to the degree in which we are interested, but since continuously recording meteorographs are now obsolete, no better equipment is in common use. The average ascensional rate has been estimated at 500 ft/min, which would yield from one to three temperature readings every 1000 ft up to 10,000 ft.[39] This is equivalent to a

vertical sampling at intervals of about 100–300 m or to a range of about 12–37 mb at levels near the average top of the inversion. The effect of this rapid ascension is to smooth the sounding by missing exact maxima and minima in the vertical temperature profile.

The computation sheets show some corrections which appear to have been entered after the observations were transmitted, but to our knowledge the data have not been subjected to systematic review and checking, as is normally done with U.S. Weather Bureau radiosonde data. By using the original computation form, however, all the errors which frequently appear due to conversion of units, to coding, to radio and teletype transmission and to decoding and replotting, have been avoided.

One other source of error, revealed by the handwriting on the original records, is the extent to which personal judgment determines which levels are critical in encoding an essentially smooth curve. The regular alternation of complex with simple ascents, in apparent agreement with the changes in handwriting, may also be due to diurnal or other real changes in the atmosphere. A detailed study on a diurnal basis, however, has not been attempted here.

B. Data reduction

1. *Processing method*—Since modern radiosonde data are available only in terms of pressure rather than height, it was decided to make this analysis using the parameter, P, which is more convenient both with respect to processing and future comparison of new data. Heights have been computed only for annual mean soundings for each of 12 types.

Up to seven levels of significant change in lapse rate were selected to describe the inversion pattern below 700 mb. For each, P_1, T_1 and the differences ΔP_1, ΔT_1 for the corresponding layers, were tabulated. When we found that correspondingly numbered levels changed greatly in height, it was necessary to simplify the sounding into three relative layers, A, B, C, which described the main features of all ascents. Only in this way could corresponding portions of a sounding be compared easily. Details on the system of classification are given in Section C below. The information entered on to IBM punch cards included the values of ΔP and ΔT for the A, B, C layers and for the seven layers, an ABC type number, and the available surface data.

The IBM cards were then tabulated to give frequencies of one variable in terms of fixed classes of another. As there were 26 fields, only a few of the 650 combinations could be examined. Some 75 pairs were selected originally, but cost, machine time and processing limitations reduced the number to some 50 tabulations. These were condensed into the frequency tabulations which are reproduced in *Tables 6–62*. Monthly analyses were made only for comparisons of ABC types and each of 11 elements. All other tables show the distribution over the entire period.

To serve as an indication of the complexity and rapid changeability of the vertical temperature field, the complete series of twice daily ascents is presented graphically for each of four individual months representative of the seasons. Prevailing surface weather for the 12 hr between ascents was obtained from the microfilm of surface records.

Suppl.
Part.I
1956

2. *Frequency distributions*—The choice of class intervals by which elements were machine tabulated and later condensed into frequency distribution tables was determined by the significance of certain values and ranges of values, by the degree of useful concentrations and by time limitations. In some instances it was desirable to use unequal classes for frequency distributions because certain values had special interest.

C. The classification of radiosonde observations

1. *Seven-level typing method*—To classify the structure of ascents, a concise classification code to describe the sequence of sign of lapse rates was first attempted by applying a "seven-level" type to each sounding. For each level of significant change in lapse rate, as determined visually from enlargements of the microfilm, a figure 1, 2 or 3 was assigned to describe respectively a decrease, no change or an increase in temperature with height in the layer above that level. A maximum of seven levels was used to describe the ascent up to 700 mb or the level where temperature decreased normally with height, whichever occurred first. Each code group was considered to be preceded by a decimal and only as many digits were used as needed. Isolated high-level inversions near 700 mb or above, which were due to fronts or subsidence, were ignored, although all significant lower inversions below 700 mb were described, regardless of their cause. Isothermal conditions (code 2) were interpreted to include almost isothermal layers of about ± 1°C/100 mb, depending upon the location of the layer. Thus ± 1°C from 800 to 700 mb would be considered isothermal, while ± 1°C from 1000 to 900 mb probably would not, because the true lapse rate in the latter instance is much greater than in the former.

For two reasons, however, such a system was found unsuitable. First, the level subscripts T_1, T_2, T_3 did not always refer to layers comparable with respect to inversion characteristics. Thus, levels with the same subscript might in one instance be near the surface, at the base of a large inversion, and in another case be near 700 mb at the top of a deep but weak inversion layer (e.g. in *Fig. 2*, note the second significant level in types 032·0 and 032·2).

Second, the exact determinations of "significant change", and of the top of the inversion layer where the temperature first began to "decrease normally", were in need of a more objective method than visual inspection, due to the finely graduated spectrum of lapse-rate changes.

2. *ABC typing method*—The first difficulty above was remedied by the division of all ascents into three possible *layers*, *A*, *B* and *C*, such that their upper *levels*, *A*, *B* and *C* were defined as:

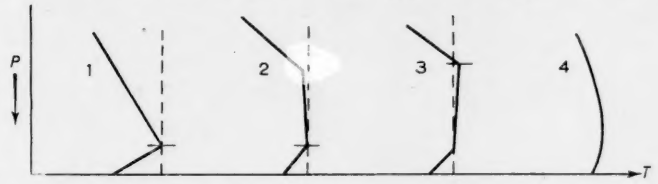
A: The top of the mixing, or cold layer, in which the temperature decreases or remains constant with height. Layer *A* must start at the surface and have a *B* or *C* layer above.

B: The top of the pronounced inversion layer. *B* layer may occur at the surface or above an *A* layer.

C: The top of a layer with small lapse rate. *C* layer is generally above an *A* or *B* layer, but may occur alone.

By restricting comparisons of levels to the same *A*, *B* or *C* level more valid results were available.

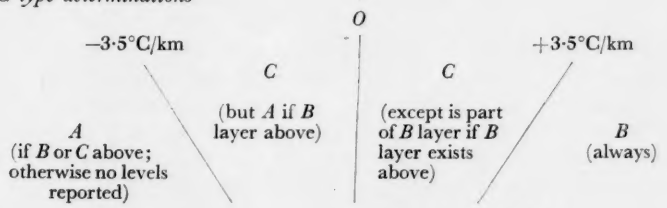
The second deficiency above required the application of a fixed measurement to determine the top of a poorly marked inversion. If the level of maximum temperature were to serve as the top of the inversion, then, as sketched below, great differences could occur in ΔP for negligible changes in ΔT:



Case (1) is clearly defined, but (2) and (3) are essentially the same ascent despite the great difference in height of the inversion top. For a gradual change in lapse rate as in (4), levels of significant change as well as the "top" cannot be determined uniquely. As such isothermal and other layers with small lapse rates are of importance with respect to radiation conditions in the lower troposphere, it was decided to include as part of the "bulge" all layers in which the lapse rate below 700 mb was less than $\pm 3{\cdot}5$°C/km. This value was chosen when measurement of many individual and mean monthly ascents revealed that the level where this lapse rate was reached seemed the most appropriate division of the stable surface layer from the relatively large lapse rates of the upper troposphere. It is interesting to find that FLOHN[14] used 3°C/km as an upper limit for the top of the main inversion layer.

Use of a transparent overlay of the $\pm 3{\cdot}5$°C/km critical lapse rate to measure the original ascent curves made it possible to determine objectively the three critical levels *A*, *B* and *C* as follows:

ABC type determinations



Coding of ABC.D types

ABC

0	missing
1	temperature decreases with increasing height
2	temperature constant or quasi-constant ± 1°C/100 mb
3	temperature increases with height

·D

0	simple layers
1	multiple *A* layer
2	multiple *B* layer
3	multiple *C* layer
4	multiple *A* and *B* layers
5	multiple *B* and *C* layers
6	multiple *A* and *C* layers
7	multiple *A*, *B* and *C* layers

Suppl. Part. I 1956

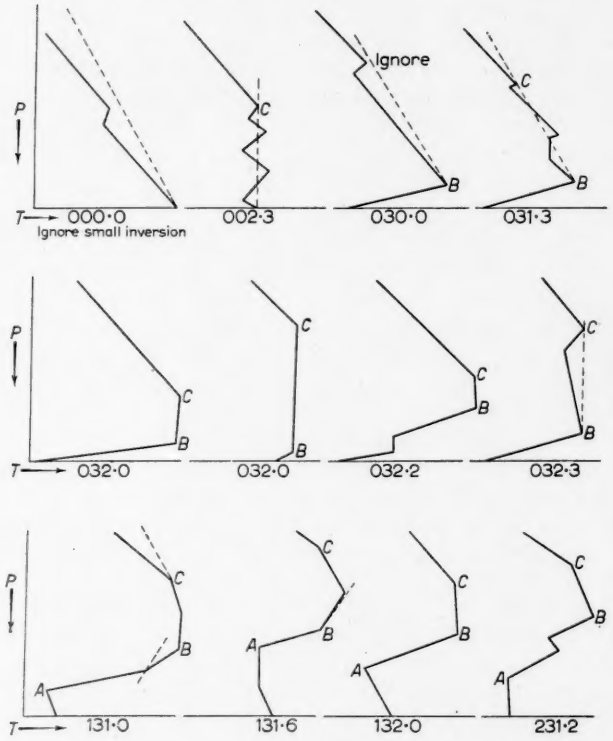


Fig. 2. Examples of ABC.D typing and coding

There are 24 possible types since *A* is allowed to have the values 0, 1, 2; *B* may be 0 or 3; and *C* may be 0, 1, 2, 3. See *Fig. 2* for examples of *ABC.D* coding. It is realized that the thickness of a millibar increases 50 per cent between 1020 and 700 mb, but the use of a lapse rate in terms of height, on a pressure scale, introduces very little error when applied to a log *P*–*T* chart such as was used, since the log *P* scale corresponds very closely to a linear height scale. In fact, the error is not any greater than the variation in slope of the adiabats over the same range of pressures and temperatures. The theoretical effect of this error would tend to raise slightly the height of *B* and *C* levels in summer and to lower them in winter. But as the template was based on transition season data, it is not felt much error is due to this from one season to another (i.e. from the warm to cold areas of the diagram).

It is interesting to note that KOHL[22] has devised a nomogram for the purpose of finding exact lapse rates in height units from certain adiabatic charts.

Since almost all of the quantitative results of this paper are given in terms of *A*, *B* and *C* layers, some further details concerning their selection are included here:

(1) Any one or combination of layers may be missing.

(2) If a portion of the ascent curve coincides with the template limit, that

level of the layer was used which showed the greater change in lapse rate from the neighboring layer.

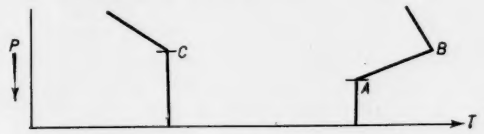
(3) Complicated ascents, with multiple inversions, termed "zig-zags", were simplified to a mean curve, and the mean curve classified as usual. Code figure .*D* was used to designate the layers which were complex.

(4) An *A* level exists only below a *B* level or if the curve falls to the extreme left of the template. *A* levels may exist with a lapse rate as great as +1°C/100 mb if below a *B* level. However, since *A* level is called "mixing layer" and implies decrease with height, the layer is considered part of the *B* layer if the increase is $>$ +1°C/100 mb.

(5) The *B* level must always lie to the extreme right of the template, but lower portions of the *B* layer may lie between 0 and +3·5°C/km.

(6) The *C* level must not have an essential inversion above it.

(7) An isothermal layer from the surface will be an *A* layer only if it is surmounted by a *B*; otherwise it is a *C*.



(8) If a portion of the ascent was missing or marked "doubtful data", the entire ascent was eliminated.

(9) If the ascent is everywhere to the left of the −3·5°C/km line, relatively shallow isothermal or inversion layers high up are ignored, and the ascent is counted as one without inversions.

D. R_2 as a measure of lapse-rate "intensity"

The normal lapse rate $\Delta T/\Delta H$, or as used in this paper, $\Delta T/\Delta P$ (here called R_1), is sufficient in itself to describe the *slope* of the lapse rate as measured from isothermal conditions, but it cannot distinguish between very short and very long *duration* of such lapse rates. We wish to have some convenient parameter of both the slope and size of an inversion and therefore introduce an arbitrary term $(\Delta T)^2/\Delta P$ which we denote in this paper as R_2 and propose as a measure of inversion "*intensity*" and hence also of inversion "*importance*".

ΔT, in itself, is a good general measure of the importance of an inversion. But since each of the indicated slopes in (1) has equal ΔT value, we want to increase the relative standing of that curve with the greatest change of temperature per unit change of pressure, and do so by weighting ΔT with $\Delta T/\Delta P$. The opposite situation (2) of constant $\Delta T/\Delta P$ requires a weighting factor ΔT and results in the same R_2 term. The essential basis for selection of ΔT, rather than ΔP, as the weighting factor is that we are more interested

in those inversions with large temperature, rather than large pressure change. Also, by using ΔT we can avoid the additional error due to the changing thickness of a millibar with height. Indeed, the use of a pressure scale for such studies can be tolerated only in view of the impracticability of computing many thousands of heights. The results of our measurements are given in III.

E. Interpretation precautions

We summarize here some of the weaknesses of the data and of our method which should be considered in interpreting the results which follow.

(1) Observational errors which have remained undetected.

(2) The bias of the monthly distribution of the sample, when discussing monthly or seasonal frequencies of variables. There is also probably a diurnal bias which, however, does not enter into our results since no diurnal studies have been made.

(3) ABC level determinations. Despite a template method, there are borderline cases in which a choice exists. Although all of the rare cases have been verified carefully to insure they are not accidentally misclassified, and all the typing was double-checked, the distribution of types might be slightly altered due to errors. More likely is the chance of wrongly positioning a level, thus affecting the values of the pressure and temperature, although the type may be unaffected. Our method has the effect of smoothing out complex layers. For this reason, the *B* layer cannot be associated with either the lower or main inversion layer, since the template divided the sublayers at the 3·5°C/km slope, rather than at the level of greatest change in slope of the sublayers.

(4) The effect of using unequal classes in the frequency distributions must be remembered.

(5) A slight seasonal tendency exists for our measurements to raise the height of the *B* and *C* layers in summer, and to lower them in winter, due to the constant height scale template.

III. Results

A. Characteristics of types

1. *Classification of types by structure and season*—Of 24 types which are possible with our permitted values of *A*, *B* and *C*, a total of 21 types was observed. Of these, 13 types including 000 (no inversion), account for 96 per cent of the 1,097 observations. Type 031 was the most frequent, with 28 per cent of all observations. *Table 3* gives the detailed distribution of all 21 types by month. From *Table 3A*, we see that the seven most frequent types describe 84 per cent of all ascents. In *Tables 12–32*, which compare various elements by ABC type, all 21 existing types are listed. However, in the remaining tables only the 13 most frequent types are used.

The types may be grouped into four general classes: surface inversion, inversion above a mixing layer, isothermal or quasi-isothermal, and inversionless ascents. The relative frequency pattern of these classes reflect the thermal regime of the lower troposphere for this area, as shown in *Fig. 4*. In terms of surface inversion, the months from November to April comprise

a winter; June–August, a summer; and May, September and October appear as transitional months.

We may now summarize the frequency of general types in terms of these seasons: About 85 per cent of *surface inversions* occurred in winter, 10 per cent in the transitional months and 5 per cent in summer. *Mixing layers* were distributed to show 45 per cent in summer, 30 per cent in winter and 25 per cent in the transition months. *Quasi-isothermal* types were most frequent in summer (65 per cent), while there were 30 per cent during the transition months and 5 per cent in winter. Ascents without any inversion were confined almost entirely to the summer.

Considered in terms of seasons, in *winter* the relative frequency of surface inversions (82 per cent) to mixing layer types (17 per cent) is almost five to one. Ascents without inversions or quasi-isothermal layers do not occur. In *summer*, however, mixing layers are the predominant form (55 per cent), isothermal ascents occur about 25 per cent, surface inversions only 10 per cent, and no inversions almost 10 per cent of the time. During the *transitional* months, 45 per cent of the observations show mixing layers, 40 per cent surface inversions, and 15 per cent are quasi-isothermal. It should be noted, however, that there is no exclusive type for any season. Typical summer or winter profiles do occur in the opposite season.

For the *year* as a whole, surface inversions comprise 56 per cent, mixing layers 32 per cent, quasi-isothermal 10 per cent, and inversionless ascents 2 per cent of the total observations.

The frequency of complex layers, indicated by D, is summarized in *Table 4*. In general, about half of the frequencies for each type are for "pure" *ABC.0* types, i.e. the layers have smooth, gradual rates of internal change of lapse rate. Complex sublayers appear about equally in each layer of a given type.

Fig. 3 summarizes graphically the three most frequent types for each mid-seasonal month (November is used in place of October due to the poor sample of October). *Table 66* gives similar data for all months.

2. *Features of most important types—Figs. 5a, b, c, d* present the entire data given in *Tables 1* and *2*, in order of decreasing frequency of type. The box gives the percentage of total occurrence, N, of the type by month. We will discuss the monthly variations of structure of the most frequent types of ascents, and note especially those which have the largest values of monthly and annual mean inversion magnitude (ΔT, lapse rate (R_1) and intensity (R_2). Monthly means of less than five cases are represented by dashed lines in the figures and are not considered here. The few examples of inversions in types 230–233 are ignored.

031. This is by far the most frequent type with 28 per cent of all observations. The winter months for which it is typical show little variation in mean appearance; the five warmest months show great variability but this is natural because of the type's infrequent occurrence. The annual mean is typical of the winter months.

032. The 15 per cent represented by this type follow the same pattern as 031 with respect both to monthly occurrence and magnitude of layers. This type has the greatest mean annual inversion magnitude in the B layer,

+13°C, although other types have larger mean monthly inversions. (The one observation in month 7 was omitted due to the additional length of scale required to include it.)

131. The most common summer type (14 per cent) presents the mixing layer typical of maritime ascents, although the cause here is more a turbulent disturbance of the radiation inversion than a turbulent layer under a subsidence inversion. The annual mean represents the warmest months. It is relatively frequent in spring and fall.

033. This type, typical of mid-winter, has the greatest monthly and annual mean inversion intensity. It also has the largest monthly and annual mean lapse rate. Further, if the total inversion amount in both *B* and *C*

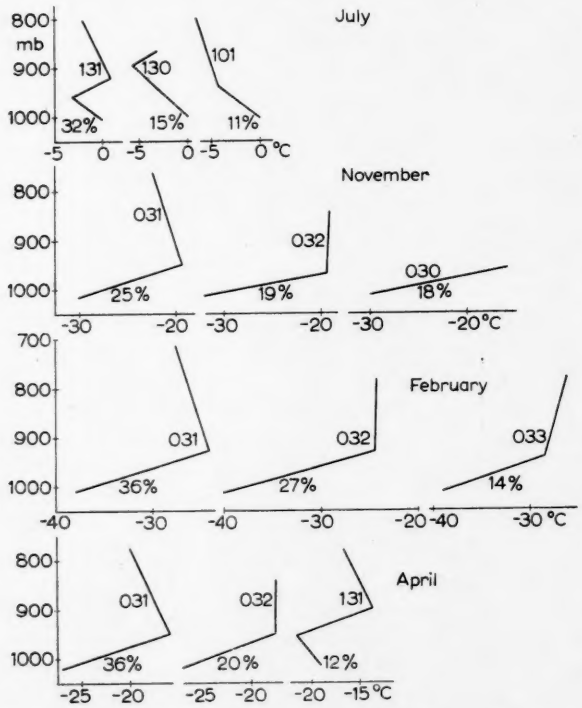


Fig. 3. Structure of most frequent inversion types for July, November, February and April

layers is considered, it has the largest annual mean magnitude as well. This type does not occur appreciably from May to October. Despite these qualifications, absolute extreme, individual inversion values are found with 031 and 032 more often than with 033.

130. This predominantly summer type also extends noticeably into both spring and fall. Its winter occurrences are erratic and often very extreme. The annual mean is typical of summer.

132. This type came most often in June, but is found throughout the year with a slight secondary peak in December. The annual mean is typical of

December. It has the least stable mean annual lapse rate in the *A* layer of all types (*Table 63*).

030. The claim to distinction here is that of having the greatest mean monthly inversion in the *B* layer (+17·6°C in March). This type is the simplest of the true inversion types. It came most often in November, and

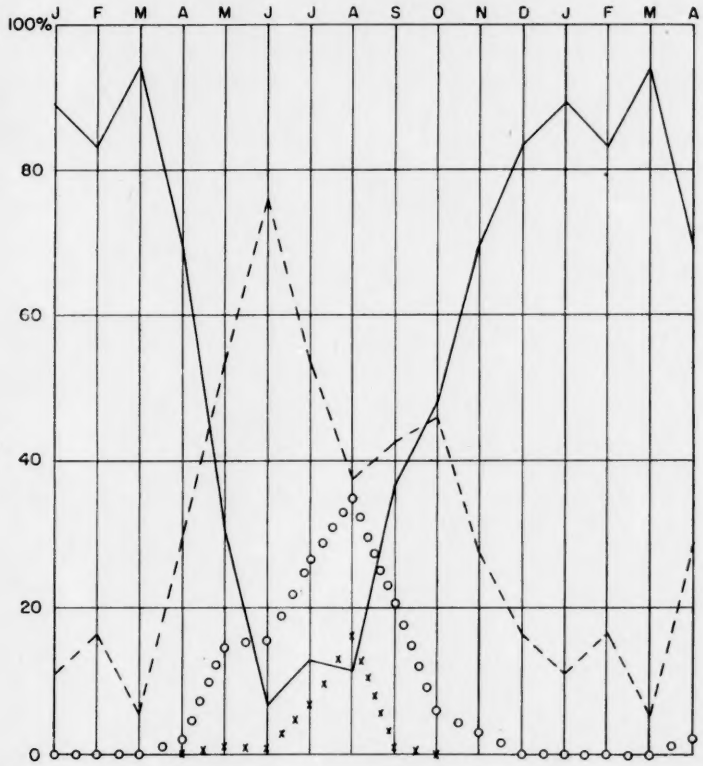


Fig. 4. Frequency of general inversion types by month

with small magnitude but surprising frequency in July. The annual mean is most typical of winter and spring.

000. Ascents without inversions or without quasi-isothermal layers occur seldom (2 per cent), and are confined to the months of May to September.

133. This rare type occurred only in spring and fall, and the few instances show a tendency for strong lapse rates in the inversion layer.

001, 101, 102, 002. All quasi-isothermal types occur most often in late summer and fall.

Suppl.
Part.I
1956

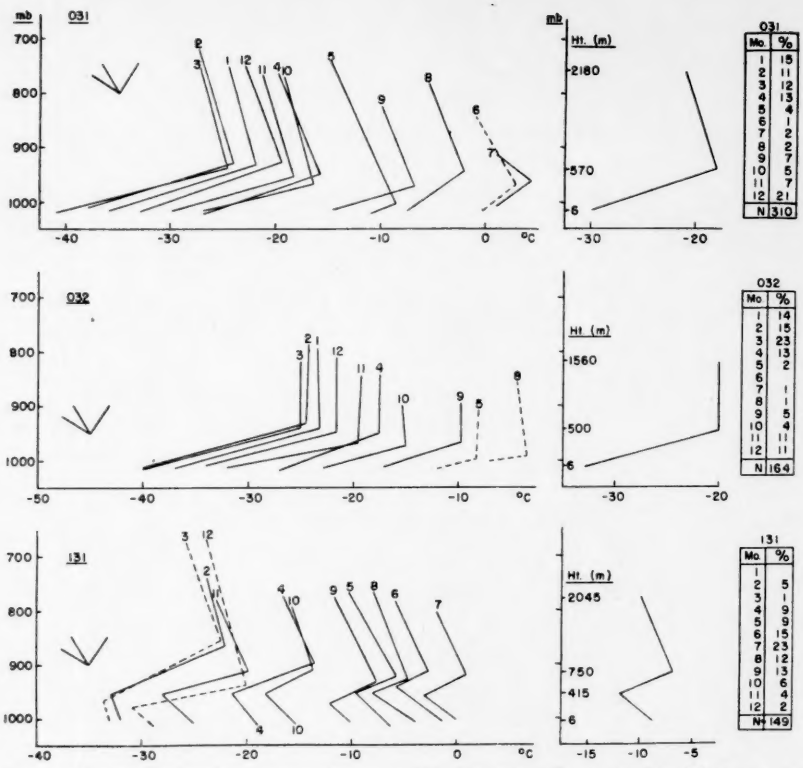


Fig. 5a. Monthly and annual mean inversion types: 031, 032, 131. The three indicated lapse rates at the left represent, respectively, −10°C/km, −3.5°C/km, and +3.5°C/km. Means of less than five ascents are shown by dashed lines

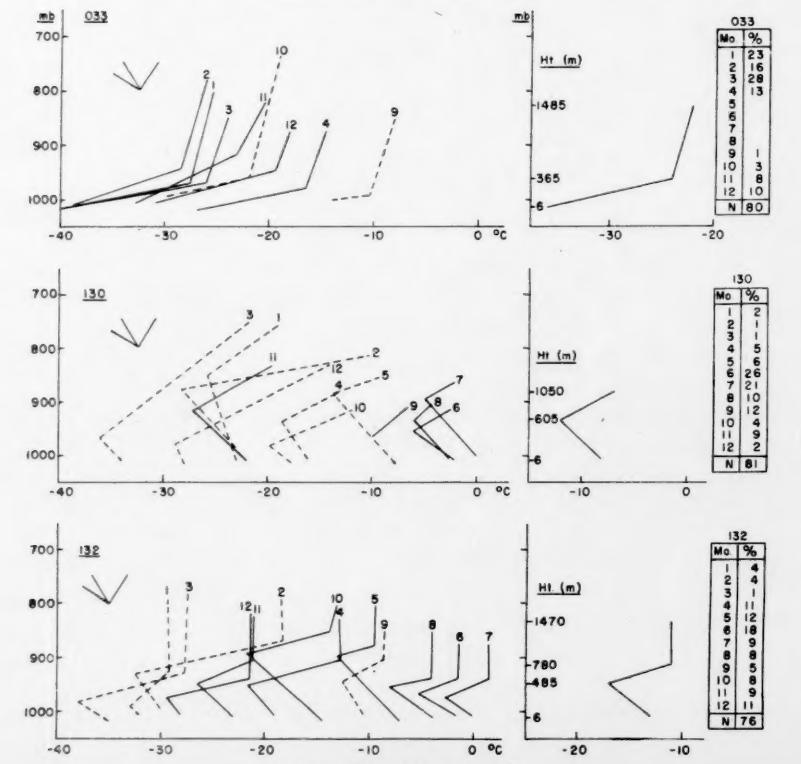


Fig. 5b. Types 033, 130, 132

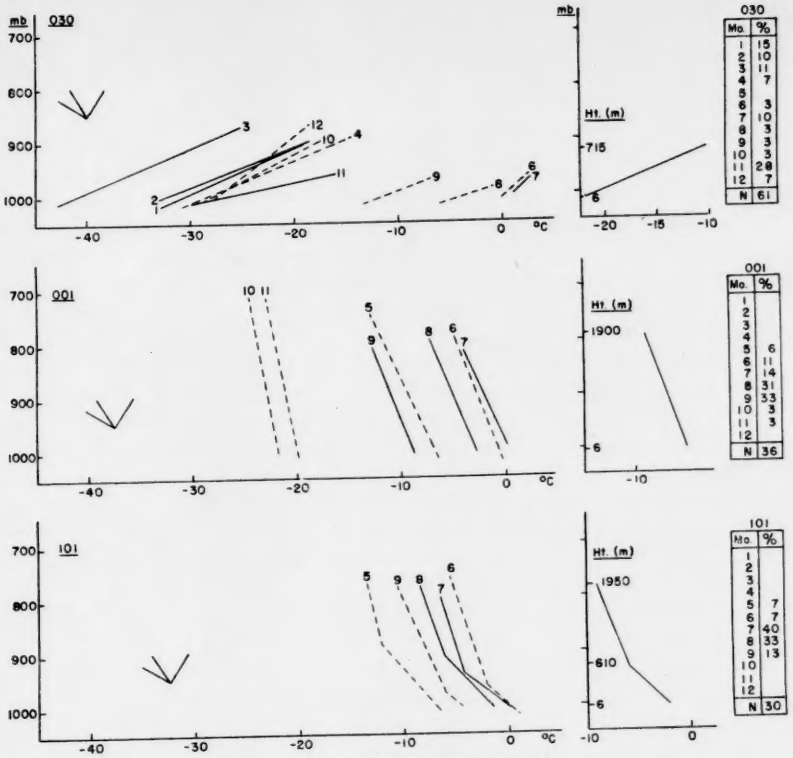


Fig. 5c. Types 030, 001, 101

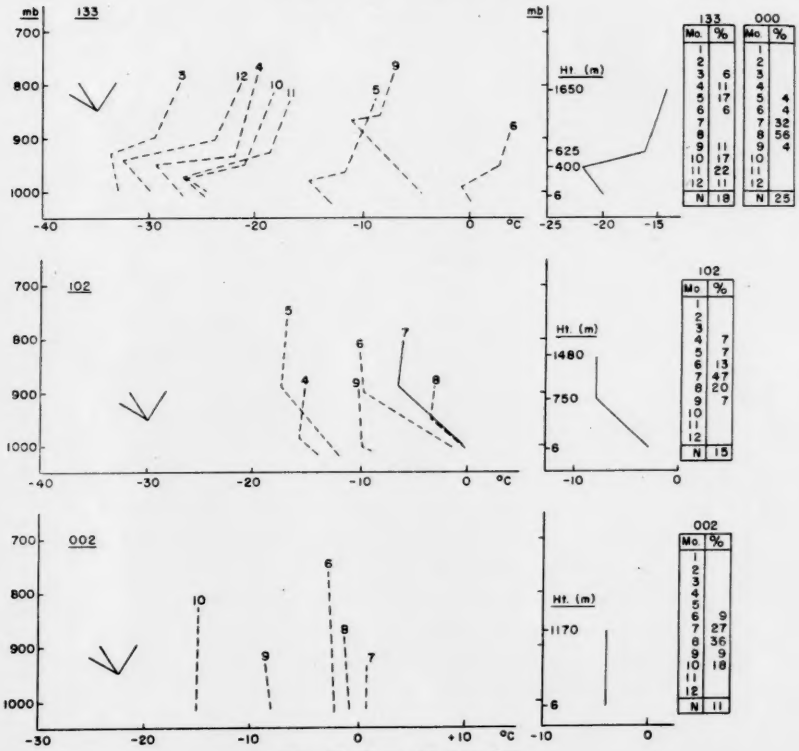


Fig. 5d. Types 133, 102, 002

Suppl.
Part. I
1956

B. Properties of the mean inversion layer

Using *Tables 1, 2, 63, 64* and *65*, we can note the maximum monthly and annual mean values of inversion magnitude, slope and intensity. In *Table 63*, $\overline{R}_1$ is given as the ratio of the means of ΔT and ΔH. It is realized that this ratio is not the mean of the individual ratios, but for practical reasons no better approximation is possible. In effect, we are simply measuring the lapse rate of the mean annual ascents shown in *Figs. 5a, b, c, d.* In *Table 64*, we again must resort to ratio of means for an approximation to a mean value of R_1, and of R_2 where the error is increased by multiplication by a mean. If we may use these ratios in a very relative manner and give due weight to frequency of type, we find the rather normal conclusions below:

1. *Inversion magnitude* (ΔT_b)—The maximum monthly mean was +17·6°C in March with type 030. When the monthly means of ΔT_b are graphed for all types, there is seen a flat summer minimum from May to August, a regular increase to a maximum near March, and then a sudden decrease from March to May. This bears a striking similarity to the reversed pattern of surface and upper-air temperature in the Arctic (cf. ref. 3, *Figs. 5–12*). This similarity is not surprising since the rapid decrease in inversion magnitude should occur with the rapid increase in insolational heating.

With respect to the annual mean, there is little difference among the four surface inversion types. Type 032 has a maximum mean of +13°C but the others all have +12°C. However, by combining inversions of the *B* and *C* layers in type 033, there is a total mean inversion magnitude of +14°C.

2. *Inversion lapse rate* (R_1)—The maximum monthly mean lapse rate is +0·294°C/mb with type 033 in March which is more representative than a single observation of +0·450°C/mb in September with the same type. There seems to be a tendency for a relative maximum in lapse rate in November for other types.

For the year as a whole, the most pronounced positive lapse rate again appears with type 033 (typical of winter), which attains +0·245°C/mb, or +33°C/km, and a least marked one with type 131 and 130 (summer types), as might be expected. It is a bit more surprising to find that the secondary maxima occur equally with 133 (equinoxes) at +26°C/km, and with 032 (winter).

3. *Inversion intensity* (R_2)—Of the four surface inversion types which have the strongest inversions, types 033 and 031 have their monthly maxima in March, while 030 and 032 have theirs in November. The values range from 3·3 to 4·1(°C)²/mb, as compared to the range of R_2 values from 0 to 32·9(°C)²/mb when all individual cases are considered. For the year as a whole, type 033 has the greatest annual mean intensity, 2·9(°C)²/mb.

To compare the use of R_1 and R_2, *Table 64* has been prepared which gives the mean value of R_1 and R_2 for surface temperatures at 1° intervals, for all *B* layers, regardless of type. We may notice that 0°C is the most frequent single temperature observed, and that 0°C the mean slope of the *B* layer (R_1) is a minimum of 0·08°C/mb.

As T_1 decreases toward winter values, $\overline{\Delta T}_b$ increases, and more rapidly than $\overline{\Delta P}_b$ increases. Thus $\overline{R}_1$ also is greatest at coldest temperatures. If we compare $\overline{R}_1$ with $\overline{R}_2$, however, we see that $\overline{R}_2$ has increased much more

rapidly due to the weighting effect of increasing $\overline{\Delta T_b}$'s. $\overline{R}_1$ by itself would not give the same relative order of importance to the inversions as $\overline{R}_2$, as can be seen by listing $\overline{R}_2$ and $\overline{R}_1$ in order by $\overline{R}_2$. Therefore $\overline{R}_1$ cannot serve the same purpose as $\overline{R}_2$, a purpose we consider desirable.

If we examine the individual values of R_1 and R_2, we also find the same lack of correspondence in order. The highest 20 values of R_2 for the entire period (ranging from 32·88 to 10·11) include the 13 highest values of R_1. The remaining 7 values of R_1 include values as low as the fifty-sixth position in the R_1 list.

In general, we would find if all inversions are compared, that small, or weak unimportant ones, and large, or intense important inversions are more appropriately rated by R_2 than R_1.

4. *Extreme inversions*—The maximum individual value of R_1 was 2·01 °C/mb and of R_2 32·88(°C)²/mb. The first case corresponded to a change of 14·1 °C in 7 mb in November (*ABC.D* type 032·0), and the second instance to a change of 17·2 °C in 9 mb in April, also with a type 032. All the 20 most extreme R_2 values occurred with a surface type inversion in winter except one which had an *ABC.D* type of 232·5 in September.

A listing of all 966 inversions for the entire period in order of R_2 shows that, in general, most "intense" inversions occur as expected in winter, although there are unusual exceptions, such as one in June with a type 130 ($R_2 = 5{\cdot}4$, $R_1 = 0{\cdot}8$). About 90 per cent of R_2 values are less than 5·0:

R_2 (°C)²/mb	Cumulative percentage	R_2 (°C)²/mb	Cumulative percentage
⩾10·0	2	5·0	11
9·0	3	4·0	17
8·0	4	3·0	27
7·0	5	2·0	41
6·0	8	1·0	63
		0·0	100

The maximum observed temperature increase in the *B* layer was 30·5 °C on 4 April 1954 when the surface temperature was −37 °C, pressure 1016 mb, wind 3 m/sec, with a type 030·2. The inversion layer was composed of several sublayers extending through 180 mb above the surface. The R_1 value is only 0·17 °C/mb but the R_2 rating is 5·17(°C)²/mb, again illustrating the better evaluation of the importance of an inversion by R_2 than R_1.

There is no doubt that more intense inversions are contained in our collection, but they occur in sublayers within our simplified *ABC* strata. If a further analysis of these data can be made, a study of the finer details of the *B* layer should be undertaken. From our definition of layers, we know only that every sublayer inversion has a lapse rate of at least +3·5 °C/km.

C. Daily sequence charts

Figs. 6a, b, c, d are presented to show, for seasonal months, the rapid variability of the vertical thermal gradient, and the relative seasonal changes in gradient, and to attempt to show the complexity of the basic problem of discussing the causes for given inversions.

Suppl. Part. I 1956

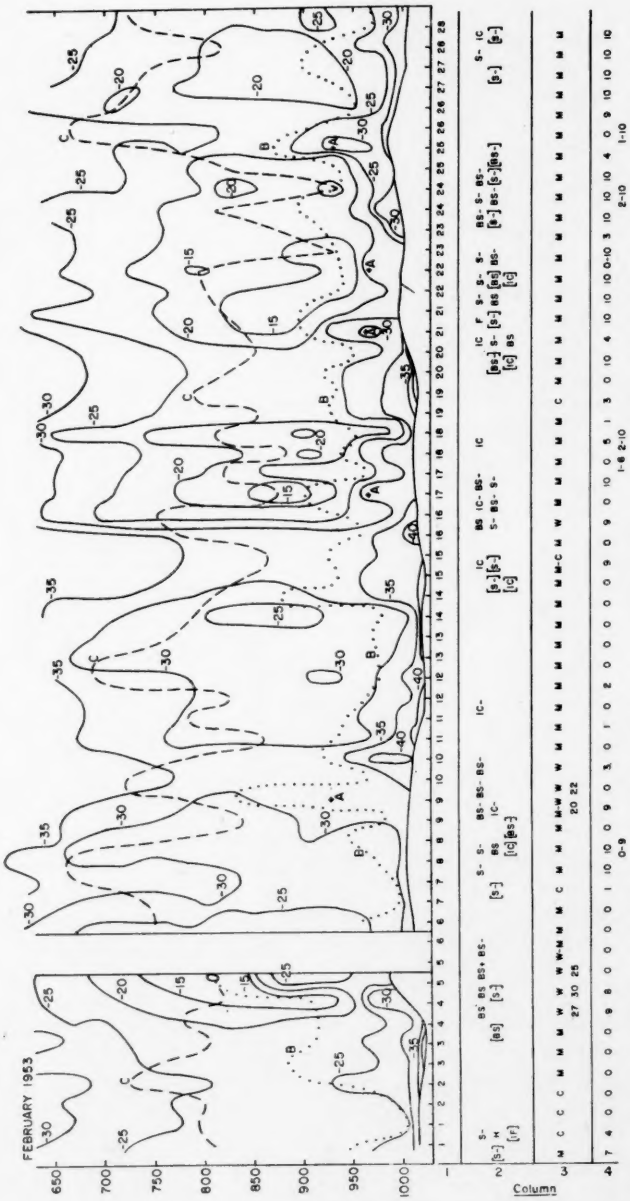


Fig. 6a. Daily sequence of inversions, February 1953

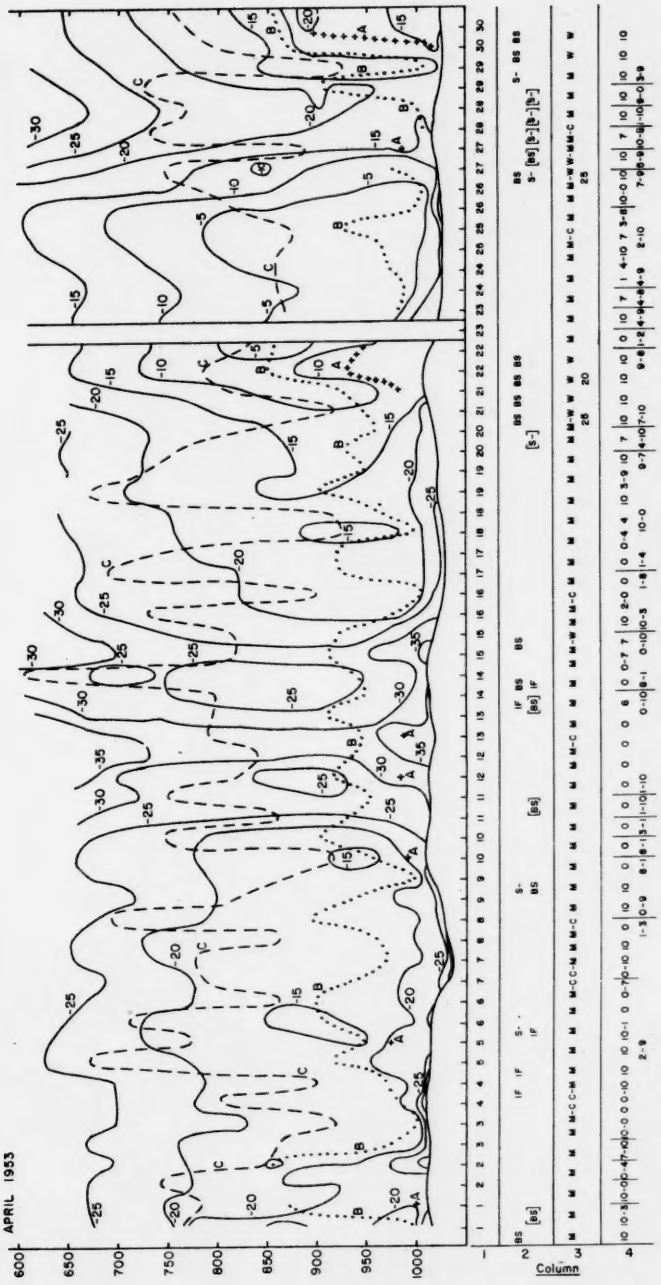


Fig. 6b. Daily sequence of inversions, April 1953

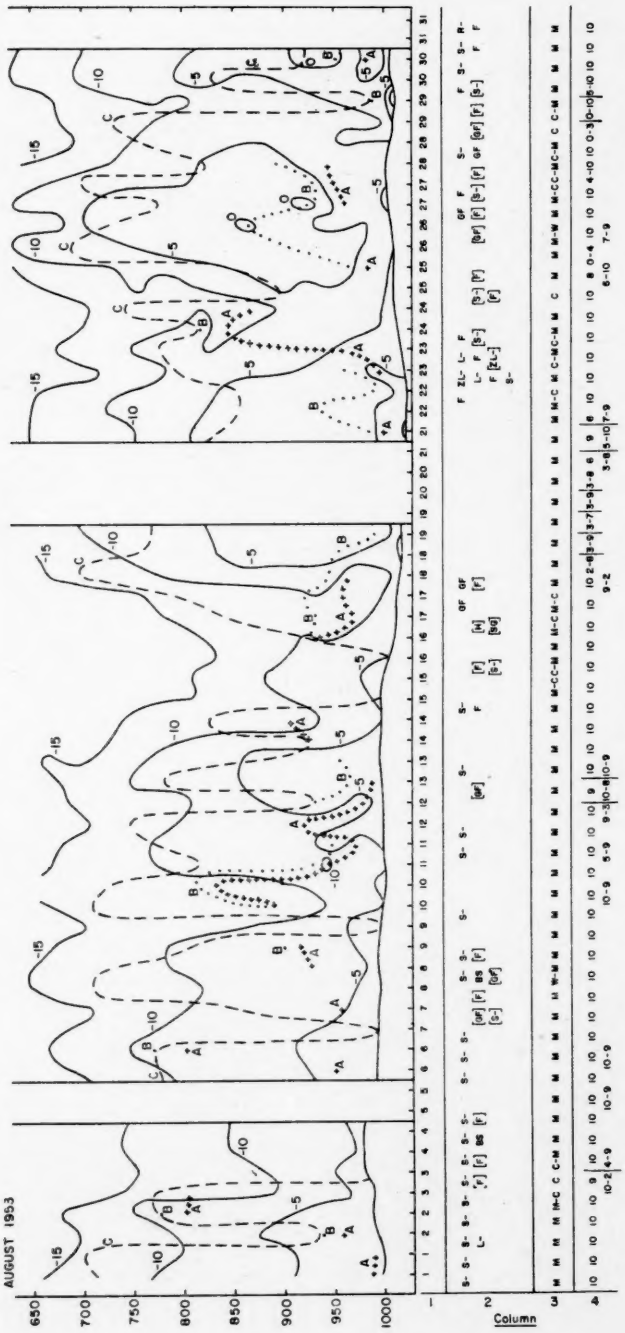


Fig. 6c. Daily sequence of inversions, August 1953

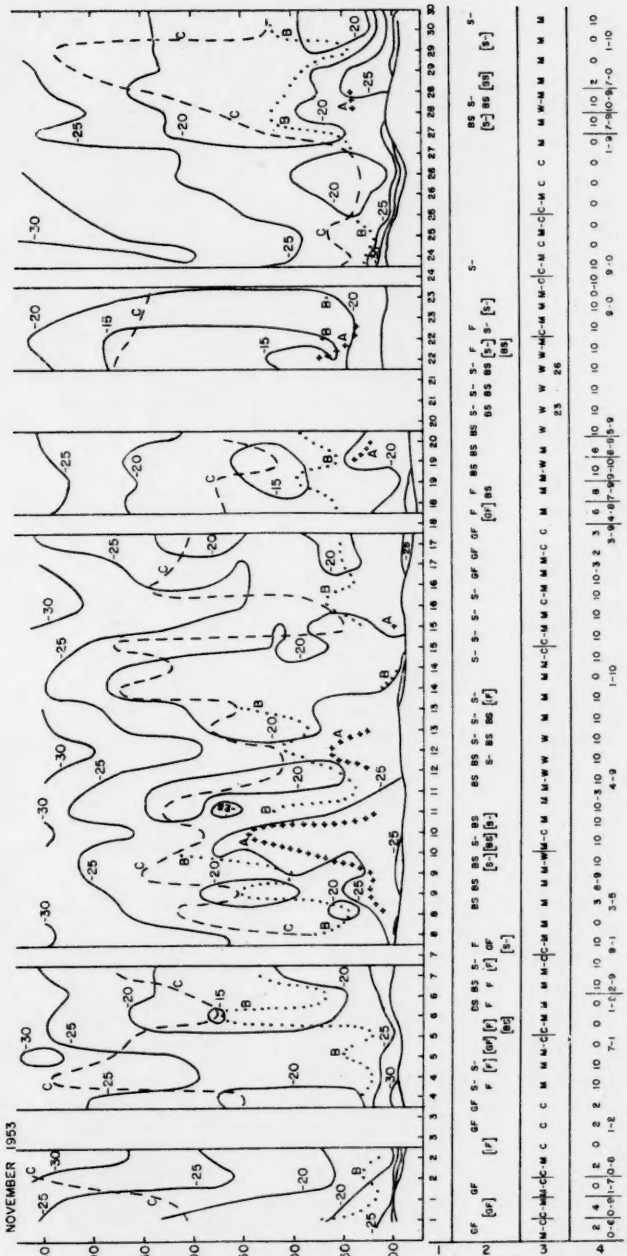


Fig. 6d. Daily sequence of inversions, November 1953

Suppl.
Part. I
1956

Explanations of symbols in figures 6a, b, c, d

———— Isotherms at 5°C intervals
— — — C level
. . . . . B level
++++ A level

The C level is continuous and at times merges with the B level.

Col. 1. Dates of twelve-hourly radiosonde ascents (G.C.T.)

Col. 2. Weather

BS blowing snow	IC ice crystals	SG snow grains
E sleet	IF ice fog	ZL freezing drizzle
F fog	L drizzle	ZR freezing rain
GF ground fog	R rain	— light
H haze	S snow	+ heavy

The symbols indicate the *prevailing* surface weather for the 12 hrs between radiosonde ascents. Those letters in brackets indicate *occasional* changes during the period.

Col. 3. Wind

Average wind speed is indicated by:
C: 0–3 knots
M: 4–19 knots
W: 20 or more knots

Gusts are shown on the second line.

Col. 4. Cloudiness

Prevailing cloudiness is given in tenths on the first line and *occasional* cloudiness on the second.

Available surface weather, averaged for the period between observations, is included. Among the features in August we note the strong tendency for weak or isothermal gradients, the relatively high frequency of the lack of inversions which is shown by the lowering to the surface of the *C* level, and the high frequency of *A* levels. *B* levels are rare. In February, the typical pattern is for intermittent, intense surface inversions and intense gradients aloft. The *C* level remains high and never reaches down to the surface. *A* levels are infrequent and isolated, while *B* levels are continuous. In November the pattern contains, as natural, features of both the above seasons. April is mainly distinguished from November by its inversion strength and continuity.

As a single example of the difficulty of trying to explain the thermal features in April from the given surface weather, we can point out the relatively clear period from 10–13 April with moderate wind, no precipitation, and relatively high pressure and low temperature. One could expect from such conditions a gradient similar to that of the period 2–5 April, but although a surface inversion exists at both times, the former has only a slight vertical gradient in comparison to the latter. The reason for this will probably also require detailed knowledge of the synoptic situation and wind field at all tropospheric levels, in addition to more complete and detailed surface weather.

D. Frequency distribution comparison of inversion parameters

The following notes indicate any general trend between the variables in *Tables 6–62*, when such a relationship is discernible. The number in parenthesis refers to table. N.R. means no relation is apparent. This is indicated visually in the tables by rectangular or circular arrangement of the frequencies. Subject to distortions of unequal classes, and bias due to sample

distribution among months, any diagonal or curved appearance of frequency tables generally indicates a relationship between variables. Abbreviations are listed in Section I B (above). A summary of ranges and most frequent values of the variables is given in *Table 65*, and means are given in *Table 1*.

1. *Surface elements*

T_1—P_1 (6): The extreme lowest pressure occurs with warm temperature and the highest pressure with coldest temperature, but no general relationship seems to exist.

W_1—T_1 (7): N.R. other than coldest temperatures occur only with the lightest winds.

W_1—RH_1 (8): Above 5 m/sec the relative humidity seems to increase as wind increases, but the relationship is poor.

W_1—P_1 (9): N.R.

RH_1—T_1 (10): As relative humidity increases so does temperature.

RH_1—P_1 (11): N.R.

2. *Surface elements by type and month*

T_1—*ABC* (12): Surface inversion types (030–033) strongly favor coldest temperatures, and mixing layer types (130–133) are most frequent at warmer temperatures. Type 000 occurs at +2 to −7°C only. The warmer the temperature, the greater the frequency of 130 and 100 types over 030 and 230 types.

W_1—*ABC* (13): N.R.

P_1—*ABC* (14): N.R.

RH_1—*ABC* (15): Summer type 130s have higher RH_1 (80–90 per cent) than winter type 030s (70–80 per cent). Low RH_1 (<40 per cent) occur only with 030s. Values >90 per cent almost never occur with 030s.

T_1—*M* (16): The sudden increase in T_1 in May, and the short summer followed by a gradual cooling to a long winter, are typical of a polar climate.

W_1—*M* (17): Although the strongest winds with which raobs were taken (12–16 m/sec^{-1}) occur in winter, there is no clear seasonal distribution evident from this table.

P_1—*M* (18): Pressure higher than 1030 mb occurred only in the cold season, but lowest values came both in summer and winter.

3. *Inversion elements by type and month*

ΔT_a—*ABC* (19): N.R.

ΔT_b—*ABC* (20): Highest values of ΔT_b occur with surface type inversions.

ΔT_c—*ABC* (21): N.R.

ΔT_a—*M* (22): Extreme maxima are found in warm months; extreme minima (and slight positive values) come in the cold months.

ΔT_b—*M* (23): Extreme maxima occur from February to April, and minima from May to August.

ΔT_c—*M* (24): N.R. except that highest positive values (+2 to +6°C) do not occur in summer.

ΔP_a—*ABC* (25): N.R.

Suppl. Part. I 1956

ΔP_b—*ABC* (26): Highest values of ΔP_b generally occur with winter types (030–033).

ΔP_c—*ABC* (27): N.R.

ΔP_a—*M* (28): The maximum thickness of the mixing layer (200 mb) occurs in the summer and fall as expected, but there is a surprising secondary occurrence in January.

ΔP_b—*M* (29): Extreme values occur mainly in February and March.

ΔP_c—*M* (30): N.R.

R_1—*ABC* (31): There is a preference for maximum inversion lapse rates to occur with surface inversion types rather than with mixing layer types. (Lapse rate is in units of °C/mb.)

R_2—*ABC* (32): Maximum inversion "intensity" occurs with surface inversion types.

R_1—*M* (33): Maximum values of inversion lapse rate generally occur in winter, but one case of $R_1 = 0{\cdot}8$ was found in June. Smallest values are preferred in summer.

R_2—*M* (34): Maximum values of inversion intensity occur in winter and smallest in summer.

R_1—T_1 (35): Maximum inversion lapse rate occurs generally with T_1 from −25 to −45°C.

4. *Relations of surface and inversion elements*

T_1—ΔT_a (36): As T_1 decreases, ΔT_a decreases.

T_1—ΔT_b (37): As T_1 decreases, ΔT_b increases.

T_1—ΔT_c (38): For 50 per cent of all cases ΔT_c is quasi-isothermal (+2 to −2°C). Largest values (+2 to +6°C) occur at coldest temperatures.

T_1—ΔP_a (39): As T_1 increases, there is a slight tendency for larger ΔP_a.

T_1—ΔP_b (40): As T_1 decreases, ΔP_b increases.

T_1—ΔP_c (41): N.R.

W_1—ΔT_a (42): As W_1 increases >11 m/sec, ΔT_a has a very slight tendency to prefer values near zero.

W_1—ΔT_b (43): N.R. except that strongest winds tend slightly to be associated with smallest ΔT_b.

W_1—ΔP_a (44): N.R.

W_1—ΔP_b (45): No clear relation, but there is a small suggestion that the stronger the wind the smaller ΔP_b.

W_1—ΔP_1 (46): N.R.

P_1—ΔT_a (47): N.R.

P_1—ΔT_b (48): N.R.

P_1—ΔT_c (49): N.R.

P_1—ΔP_a (50): N.R.

P_1—ΔP_b (51): N.R.

P_1—ΔP_c (52): N.R.

RH_1—ΔT_a (53): As humidity increases there is some suggestion that ΔT_a attains larger negative values.

RH_1—ΔT_b (54): Suggests that as RH_1 increases, ΔT_b decreases.

RH_1—ΔP_a (55): N.R.

RH_1—ΔP_b (56): N.R.

5. *Internal relations of inversion elements*

ΔT_a—ΔT_b (57): As ΔT_a becomes more negative, ΔT_b decreases.
ΔT_a—ΔP_a (58): As ΔT_a becomes more negative, ΔP_a increases.
ΔT_a—ΔP_b (59): As ΔT_a reaches large negative values, ΔP_b decreases. As ΔT_a becomes isothermal, ΔP_b increases rapidly.
ΔT_b—ΔT_c (60): N.R.
ΔT_b—ΔP_a (61): N.R.
ΔT_b—ΔP_b (62): As ΔT_b increases, ΔP_b increases.

IV. Discussion of Results

A. Causes of inversions

Among the factors responsible for inversions are radiation, advection, vertical motions and fronts. Although all of these may contribute to Arctic inversions, the chief cause is radiative cooling of the atmosphere.

The basic theory holds that the snow-surface radiating as a black body loses more radiation than it receives from the atmosphere which emits only as a gray body. If the atmosphere's temperature remained constant, then the snow-surface temperature would decrease until its emission (a function of its temperature) balanced that received from the atmosphere. However, this ideal relationship is not found in practice since a change in one variable immediately changes all the others. The circular chain of interactions does not lend itself to simple solution. Even if such continuous interdependent temperature changes could be evaluated, we find that we do not yet know what the initial conditions of atmospheric composition are, and find the effects of constantly varying, horizontal advection and vertical motions also impossible to handle quantitatively. Granted the solutions to both these problems, we have still a relatively minor one of the physical values of the absorption and emission coefficients for each of the atmospheric constituents. These in turn are not constant, but may have important dependence upon wavelength of the incoming radiation, temperature and pressure.

The last of the three major problems is under active investigation (e.g. ref. 18, 42), and in reasonable time usable results should be available. The first problem might be examined under controlled laboratory conditions, in which the net result of given initial conditions would at least be available, and probably this would yield some information on the intermediate processes. The second seems the most difficult to evaluate in practice, and our only hope is to find examples of conditions which are reasonably uniform over the period of examination and evaluate the net effects of external influences as the balance between computed and observed values of the radiation problem.

For these reasons, all radiation theory is severely limited when applied to practical work. Simplified procedures must continue to be used since the exact solution cannot yet be derived. We wish only to point out the inherent weaknesses of any conclusions which may be based on them. In many ways, general speculation may be more justifiable and useful than rigorous attempts using non-permissible assumptions. The latter may lead to spurious, but convincing and generally accepted, conclusions due to the apparent

Suppl. Part. I 1956

soundness of the reasoning or to the accuracy of the method, while the necessary assumptions are ignored.

It would seem that statistical reduction of the observations which are now available would offer the best hope of learning the actual results of these intangible processes; possibly we can derive information on the physical processes when sufficient facts are gathered in this way.

B. Comparison of results with the literature

1. WEXLER[43, 44]—These papers are the only quantitative attempts to explain how inversions form in polar air, and apparently Wexler's theory has not again been checked since he himself attempted to do so unsuccessfully in the 1941 paper. His theory postulates that as the surface temperature decreases from 0°C the inversion forms quickly until an equilibrium is reached between atmospheric and surface radiation (in a matter of hours), and the surface temperature has reached −33°C. From then on, as the surface temperature decreases, the inversion *decreases* and the thickness of an isothermal layer increases. His ideal sounding consisted of a surface inversion of infinitesimal thickness below an isothermal layer which extended up to the resumption of the original lapse rate curve. All the other highly restrictive conditions need not be enumerated here if we mention only that there is no wind at any level, the atmosphere is clear, and no advective processes exist. Certain fixed amounts of water vapor and carbon-dioxide were assumed throughout, and their absorption coefficients were constant.

Wexler attempted to check the assumed radiation curves with radiation data from Fairbanks and Fargo, N.D., and with an Elsasser diagram. He found his radiation curves did not fit the radiation observations well and it was hoped that later studies would investigate the composition of the atmosphere to see what other constituents could reduce the outgoing radiation. In a recent communication (1956), Wexler implies that his theory of radiative cooling cannot be tested since the basic conditions for which the theory is valid do not exist in reality, and he agrees that even his observations suggest that the inversion increased as the surface temperature decreased.

FLOHN,[14] from his Yakutsk data, disagreed with Wexler's theory and claimed both that ΔT_b increases as T_1 decreases and that there is no relationship between T_1 and P_c.

Using *Table 37* for all our data, not merely for type 032 which corresponds to Wexler's ideal case, we can confirm Flohn's statement, since as T_1 decreases below −10°C, ΔT_b increases. In *Table 40* we see that as T_1 lowers to values colder than −10°C, the mode value of ΔP_b increases slowly from 0–20 mb to 40–60 mb, and the upper limit of its range increases slowly from 80–100 mb to above 200 mb. In *Table 41* no relation is evident between T_1 and ΔP_c for the coarse class interval used.

This result confirms the ordinary intuitive idea that the coldest temperatures are associated with greatest inversions. We suggest as an explanation that during clear, calm weather, with a minimum of water vapor, there are ideal radiative conditions which produce the low snow surface temperature. There must be a flux of heat to the colder snow-surface from the overlying, shallow layer of warmer air. The duration of the ideal radiation

conditions is probably of more importance than the initial temperature. The longer the snow-surface radiates to space, the lower the surface temperature, and the more the heat that is withdrawn from the lowest meters of air above the surface. Presumably the minimum possible surface temperature is limited, perhaps by turbulence, to values near −70°C at sea level. The amount of downward heat flux must decrease rapidly with height, in proportion to the vertical temperature gradient, so that lowest levels cool more rapidly than upper levels in the lowest inversion layer, thus resulting in a net increase in the temperature change of the layer, with decreasing surface temperature. Also, if the downward flux is concentrated in the lowest levels, the upper limit of the isothermal layer would not be easily affected by the surface temperature.

2. SVERDRUP[36]—In his classic, clear discussion of the Arctic inversion, Sverdrup points out the differences which exist between the summer months and the spring or fall, when insolation still exists. In summer, "the free air temperature continues to rise after the temperature at the ice has reached a constant value of zero". The air near the ice cannot deviate much from 0°C and therefore in July and August there is a mean air temperature of nearly 0°C. An inversion will develop due either to warming of the air some distance from the ice by direct absorption of heat, or to warmer air advection from other regions whose lowest layer is cooled by the ice.

In spring and autumn, at below freezing temperatures, there are convection currents above the snow-surface where it is heated by radiation from the sun and sky. At noon, the surface temperatures are much higher than the air temperatures a short distance above, which result in mixing in the lowest layer.

The *Maud* observations from 1922–25 support these ideas. In winter, the inversion amounts to 7°C, in July to 6°C (but only three observations), and in June, August and September it is about 2°C.

In his summary of the "Ptarmigan" flight data, POAGE[29] confirmed this secondary July maximum for observations at 80° and 90°N, but not at 75° and 85°N.

As a corollary to the relationship above, Sverdrup expected a secondary minimum in the thickness of the mixing layer in July, although he had insufficient data to show it. Poage confirmed this in July at all latitudes.

Turning to T-3 data, *Table 23* shows that ΔT_b increases steadily from May to March, then decreases again to its May minimum. No evidence of a secondary July maximum is apparent. The higher number of cases of inversions in July is only a reflection of the distribution of the original sample. In relative numbers, the percentages of ascents with inversion layers from May to October are: 84, 83, 66, 50, 79 and 94 per cent. July with 66 per cent was thus less likely than spring or fall months to have inversions during T-3's occupation and location.

In *Table 28*, ΔP_a, the thickness of the mixing layer, has maximum values in summer but it is rather irregular.

We can only conclude that our data do not reflect a secondary maximum in either the occurrence or the magnitude of inversions in July during the two Julys of record near 86–88°N. But although the total picture is not

Suppl.
Part. I
1956

confirmed, it is interesting to note that two of the July types, 131 and 132, do show the anticipated secondary maximum in ΔT_b in July and also the secondary minimum in ΔP_a. These ascents only constitute about a third of the month's total, however, and are not able to change the opposite tendency of other types.

Fig. 5c shows that minor type 030 has a relatively high frequency in July which tends to support the idea of a relative decrease in convective activity in that month. However, other factors, such as wind, cloudiness and migratory cyclones apparently counterbalance the surface radiative conditions. *Table 17* indicates a tendency for more frequent moderate winds (3–5 m/sec) in July than in September. Complete surface observations should throw more light on the reasons why a secondary inversion maximum does not appear.

Sverdrup noted that the thickness of the mixing layer was large when the inversion had a weak lapse rate, and also that the lapse rate in the mixing layer is large when that for the inversion layer is weak. This is in agreement with our findings that the mixing layer and inversion layer act in opposition to each other (III, D.5 and IV, C.3).

3. BEDEL[1]—Bedel suggests that winter surface inversions in Greenland disappear only when the wind exceeds 10 m/sec, which he terms a necessary but not a sufficient condition. He observed 12 such cases.

Table 13 shows that five surface type inversions were observed at T-3 with winds from 12–16 m/sec. It is to be noted that the sharp decrease in observations at winds above 11 m/sec is probably mainly due to the difficulty of taking the ascent. As further data are not readily extracted from our basic material, Bedel's criterion could not be verified.

C. Summary of results

1. *The classification and frequency of inversion types*—The use of *A*, *B* and *C* levels which are suggested here permit an objective summary of inversion data and frequency by referring always to surfaces which are uniquely determined, instead of to subjectively selected levels. The associated "types" are also convenient descriptions of the inversions.

The types fall into distinctive seasonal patterns. In winter (November–April), surface inversion types predominate (82 per cent). In May, September and October, surface inversions decrease to 38 per cent, while mixing layer types become 47 per cent, and quasi-isothermal layers, 14 per cent. In summer (June–August), the frequencies are: mixing layers, 56 per cent, quasi-isothermal, 26 per cent; and no inversions, 8 per cent.

For the entire period, surface inversion types comprise 56 per cent, mixing layer 32 per cent, quasi-isothermal 10 per cent, and no inversion or no isothermal layer 2 per cent of the total observations. Type 031 is the most frequent single type (28 per cent) and is followed by 032 (15 per cent) and 131 (14 per cent).

These general results may be thought of as natural consequences of the surface thermal régime. In winter, during relatively clear and cold weather, intense surface radiation cools the air nearest the ice resulting in the common, surface or "nival" inversion. In spring and fall convection becomes

important in the lowest few hundred meters and causes limited convective and turbulent mixing. In addition, cyclonic storms increase the frequency of strong and moderate winds, thus creating additional turbulence. In summer, these same factors become more frequent, resulting in the maximum frequency at that time, of mixing layer or isothermal types, depending on how well the mixing has succeeded in destroying the normal inversion. The presence of distant cyclonic activity in summer can also help account for the upper inversions at this time of year by means of warm air advection over the near freezing surface layer.

2. *Relations of surface weather and types*—The primary result is that both the inversion thickness and the temperature warming increase as the surface temperature decreases. This supports Flohn's findings and agrees with the intuitive idea that the coldest surface temperatures create the strongest surface inversions.

It is also found that, for 50 per cent of all cases, the main (C) layer varies ± 2°C from 0°C. But when the surface temperature is coldest, the C layer attains its highest values ($+2$ to $+6$°C). This appears natural when one realizes that the C layer is usually the transition zone between the inversion B layer below it with positive lapse rate, and the region of large negative lapse rate above it. However, when the surface is coldest, the inversion layer is most intense and thick as we have just mentioned, so the C layer more often becomes part of a weak inversion region itself.

The coincidence of extreme, low, surface pressure with warm temperature, and of highest pressure with coldest temperature, is explainable in terms of summer cyclones and winter anticyclones. The former advect warm air from southern latitudes and also, through turbulence, destroy any possible surface inversion; the latter are typically associated with clear, calm and cold radiation weather.

As surface temperature decreases to -40°C, the mixing layer vanishes. Thus, in the coldest month, March, only six instances of an A layer occur. The total annual frequency of such layers is 409.

3. *Internal relations of inversion layers*—We may condense the apparent relations as follows:

1. As ΔT_a increases, ΔT_b decreases.
2. As ΔT_a increases, ΔP_a increases.
3. As ΔT_a increases, ΔP_b decreases, and as ΔT_a becomes isothermal, ΔP_b increases rapidly.
4. As ΔT_b increases, ΔP_b increases.

Thus, as the temperature change in the mixing layer increases, that in the inversion layer decreases; the thickness of the mixing layer increases; and the thickness of the inversion layer decreases. Further, as the change of temperature in the mixing layer decreases to zero, the inversion layer thickness increases rapidly.

This shows that the mixing layer and inversion layer represent opposite conditions, and that as one thrives, the other declines. Association with turbulence and radiation, respectively, serves as a convenient, if over-simplified, explanation.

The temperature change and pressure change of the mixing layer increase together and decrease together; similarly for the inversion layer.

4. *Comparison with the literature*—The conclusions of Wexler's cooling theory are not verified due to the artificial conditions to which the theory applies.

Sverdrup's expectation of a secondary maximum of ΔT_b in July, partly confirmed by Poage, was not verified in general, although for restricted types of inversions it was observed.

Bedel's necessary but not sufficient condition for the disappearance of winter surface inversions (winds over 10 m/sec) could not be verified as our data were given no sequential analysis.

5. *Miscellaneous results*—(*a*) We find that a new unit $R_2 = (\Delta T)^2/\Delta P$ is a useful combined measure of lapse rate and duration of lapse rate. It thus offers a convenient unit for inversion "intensity". It is shown that R_2 and lapse rate do not yield the same relative evaluation of inversions. Very small unimportant, and very large important inversions are better distinguished by R_2 than by lapse rate. In terms of this unit, the maximum mean monthly and annual intensities for surface inversions was $3(°C)^2$/mb compared with an absolute range from 0 to $33(°C)^2$/mb for individual inversions.

(*b*) The maximum mean monthly value of ΔT_b, the thermal magnitude of the inversion, occurs in March with type 030 and reaches +17·6°C. The mean annual value has a maximum with 032 (+13°C) with regard to the *B* layer only. If we consider the combined inversions of all layers, type 033 has +14°C for the yearly mean.

(*c*) The most pronounced mean annual lapse rate in the inversion layer comes in type 033 (+33°C/km) and it is least developed with type 130 (+11°C/km).

D. Suggestions for future work with the data

With the wealth of information still hidden in the material, we suggest the following topics be investigated:

(1) Examine the characteristics of the sublayer components of *A*, *B* and *C* layers, especially with reference to the *B* layer's most intense inversion components and the *A* layer's small super-adiabatic layers near the surface.

(2) Compare the complete surface weather observations and the upper winds with the magnitudes of *A*, *B* and *C* layer parameters.

(3) Prepare probability diagrams for selected combinations and sequences of surface weather, versus *ABC* types and characteristics.

(4) Attempt sequential analysis to discover any possible preliminary indications of change in the vertical thermal gradients.

(5) Prepare detailed synoptic histories of examples of maximum intensity (R_2) and other inversion singularities.

(6) Apply our *ABC* method to the nearest land station data for comparison with T-3, and test its usefulness at other latitudes. Eventually, a planetary mapping of inversion characteristics should be prepared. Like the tropopause, variations of the lower tropospheric inversion reflect the general radiational, synoptic and dynamic features of the atmosphere.

Acknowledgments

The very large quantity of material (1100 observations each containing 30 parameters) was reduced in a relatively short time only through the intensive application on different occasions of five part-time student assistants employed by the Arctic Meteorology Research Group, by the use of IBM punch-card techniques and especially by the reliable and assiduous help of Miss Pamela Russell and my wife. To all these I extend my grateful acknowledgment and appreciation for a long, tedious task well done.

The research reported in this document has been sponsored by the Geophysics Research Directorate of the Air Force Cambridge Research Center, Air Research and Development Command, under Contract No. AF 19(604)-1141.

References

[1] Bedel, B. *Les Observations Météorologiques de la Station Française du Groenland; Conditions Atmosphériques en altitude du 17 Septembre 1949 au Août 1951;* Paris, 1954

[2] Bellamy, J. C. "Non-redundant Processing of Continuous Data" *Cook Technical Review.* Vol. I, No. 2, pp. 117–142, 1954

[3] Belmont, A. D. "Summary of Ice Island T-3 Upper Air Temperature Data, 1952–53, and Comparison with Other Polar Station Data" *Scientific Report No. 4;* University of California, Los Angeles, pp. 30–35, 1954

[4] Brocks, K. „Die Höhenabhängigkeit der Lufttemperatur in der nächtlichen Inversion" *Met. Rund.* Vol. 2, Nos. 5–6, pp. 159–167, 1949

[5] Cook Research Laboratories. "Study of the Usefulness of Unitary Differential Notation for Storing and Utilizing Meteorological Data" *Scientific Report* SR 62–1, Skokie, Ill., 1955

[6] Cramér, H. *The Elements of Probability Theory,* pp. 166–170; New York, 1955

[7] Croxton, F. E. and Cowden, D. J. *Applied General Statistics,* pp. 177–180; New York, 1950

[8] Dibner, V. D. *The Origin of the Floating Ice-islands.* (In Russian.) Translated by E. R. Hope as Canada, Defence Research Board, T 176 R, 1955

[9] Ekhart, E. „Über Inversionen in den Alpen" *Met. Rund.* Vol. 2, No. 5–6, pp. 153–159, 1949

[10] Evganov, N. "Future Flights in the Arctic" (In Russian.) *Met. Vestnik.* Vol. 38, No. 4, pp. 110–114, 1928

[11] Evganov, N. "Les résultats des observations aérologiques par des levers de cerf-volant sur le navire hydrographique 'Taimyr', faites en 1193-15"; Leningrad, 1931. (Reviewed by C. E. P. Brooks: "The Vertical Temperature Gradient in the Arctic" *Met. Mag.* Vol. 66, No. 791, pp. 267–268.)

[12] Ficker, H. von. „Die Passatinversion" *Veroff. des Met. Inst. der Universität Berlin* Vol. 1, No. 4, 1936

[13] Flohn, H. „Zum Klima der freien Atmosphäre über Sibirien. II. Die regionale winterliche Inversion" *Met. Rund.* Vol. 1, No. 1, pp. 1–5, 1947

Suppl.
Part. I
1956

[14] FLOHN, H. „Zur Aerologie der Polargebiete" *Met. Rund.* Vol. 5, pp. 81–87 and 121–128, 1952

[15] GERMANY, DEUTSCHE WETTERDIENST. *Täglicher Wetterbericht;* Bad Kissingen, 1952–54

[16] GUTERMANN, I. G. "Temperature Conditions in the Vertical at Franz Josef Land" *Meteorologiia i Gidrologiia* No. 6, pp. 56–65, 1938. (English summary by R. G. Stone in *Bull. Amer. met. Soc.* Vol. 19 (10), pp. 433–436; French review by V. Mironovitch in *La Météorologie*, pp. 74–75, 1939

[17] HANN, J. „Über die Erscheinung der Wärmezunahme mit der Höhe in den Wintermonaten" *Met. Z.* Vol. 5, pp. 513–521, 1870

[18] HERGESELL, H. „Die Erforschung der freien Atmosphäre über dem Polarmeer" *Beitr. Phys. frei. Atmos.* Vol. 2, No. 3, pp. 96–98, 1906

[19] HOWARD, J. N., BURCH, D. L. and WILLIAMS, D. "Near-Infrared Transmission through Synthetic Atmospheres" *U.S. Air Force Cambridge Research Center, Geophysical Research Papers No. 40;* Cambridge, Mass., 1955

[20] KERNER, A. „Die Entstehung relative höher Lufttemperaturen in der Mittelhöhe der Talbecken der Alpen im Spätherbst und Winter" *Met. Z.* Vol. 11, pp. 1–13, 1878

[21] KOENIG, L. S., *et al.* "Arctic Ice Islands" *Arctic* Vol. 5, No. 2, pp. 66–103, 1952

[22] KOHL, G. „Ein Gerät zur Bestimmung des vertikalen Temperaturgradienten aerologischer Aufstiege" *Met. Z.* Vol. 7, No. 1, pp. 24–27, 1953

[23] KONDYRIN, V. V. "Temperature in the Lower Layers of the Atmosphere According to Kite and Aeroplane Soundings at the Polar Station on Cape Schmidt in 1935–36" (In Russian.) *Problemy Arktiki*, Leningrad, No. 2: pp. 75–88, 1938. (Abstract in *Met. Abstracts* 7A-79)

[24] MEYER, H. K. „Das Wetter in der Nähe des Nordpols" *Met. Rund.* Vol. 8, No. 1/2, pp. 35–38, and 7/8, pp. 117–122, 1955

[25] NEIBURGER, M. "Inversion of Temperature" *Encyclopaedia Britannica*, 14th ed., 1956 revised printing, pp. 548–549

[26] PENNDORF, R. „Aerologische Studien. I. Die mittleren Temperaturverhältnisse der freien Atmosphäre am Rande des antarktischen Kontinents". *Met. Z.* Vol. 60, pp. 201–204, 1943

[27] PETTERSSEN, S. *Weather Analysis and Forecasting*, pp. 100–101; New York, 1940

[28] PIAZZI-SMYTH. "Astronomical Expedition on the Peak of Teneriffe" *Trans. roy. Soc. Lond.* V, Vol. 148, pp. 465–533, 1858

[29] POAGE, W. "The Dropsonde Record from Alaska to the North Pole" *Scientific Report* No. 2, Arctic Meteorology Research, Contract AF 19 (122)-228, University of California, Los Angeles, 1954

[30] PÔNE, R. "Essai de classification des sondages d'aprés leur structure" *J. sci. Mét.* Vol. 4, No. 15, pp. 87–98, 1952

[31] SCHNEIDER-CARIUS, K. „Der Schichtenbau der Troposphäre" *Met. Rund.* Vol. 1, pp. 79–83, 1947

[32] ——. „Der Aufbau der Grundschicht im mitteleuropäischen Klimagebiet" *Met. Rund.* Vol. 1, No. 7–8, pp. 228–231, 1948

[33] ——. „Der Inversionstyp der Grundschicht" *Met. Rund.* Vol. 1, Nos. 7–8, pp. 226–228, 1948

[34] SHAW, J. B. "Vertical Temperature Gradient in the First 2000 ft" *Met. Mag.* Vol. 84, No. 998, pp. 233–241, 1955

[35] SVERDRUP, H. U. "The North Polar Cover of Cold Air. Preliminary Results from the 'Maud' Expedition" *Mon. Weath. Rev.* Vol. 53, No. 11, pp. 471–475, 1925

[36] ——. *The Norwegian North Polar Expedition with the "Maud". Scientific Results* Vol. 2; Part 1: Meteorology, discussion, pp. 1–40; Bergen, 1933

[37] TALJAARD, J. J. "Stable Stratification in the Atmosphere over South Africa" *Notos.* Vol. 4, No. 3, pp. 217–230, 1955

[38] TAUBER, G. M. "Temperature Gradients in Various Air Masses over Dickson Island" (In Russian.) *Meteorologiia i Gidrologiia* No. 9–10, pp. 88–99, 1938. (English summary by R. G. Stone in *Bull. Amer. met. Soc.* Vol. 20, pp. 376–378, 1939)

[39] U.S. AIR FORCE, Air Weather Service, Operations Requirements Division. Personal Communication, 8 March 1956

[40] U.S. WEATHER BUREAU. *Manual of Radiosonde Observations.* Circular P, 6th ed., pp. 82–96, 1950

[41] UNIVERSITY OF CALIFORNIA, Department of Meteorology. "Meteorological Observations from the Cruise of the 'Crest' and the 'Horizon', 1949 and 1950" *Technical Report No. 1, Subtropical Pacific Meteorology Project;* Los Angeles, 1952

[42] WATANABE, K., ZELIKOFF, M. and INN, E. "Absorption Coefficients of Several Atmospheric Gases" *U.S. Air Force Cambridge Research Centre, Geaphysical Research papers* No. 21; Cambridge, Mass., 1953

[43] WEXLER, H. "Cooling in the Lower Atmosphere and the Structure of Polar Continental Air" *Mon. Weath. Rev.* Vol. 64, No. 4, pp. 122–136, 1936

[44] ——. "Observations of Nocturnal Radiation at Fairbanks, Alaska, and Fargo, North Dakota" *Mon. Weath. Rev.* Suppl. 46, 1941

[45] ZUBOV, N. "Arctic Ice Islands and How They Drift" (In Russian.) *Priroda,* No. 2, pp. 37–45, 1955. (Translated by E. R. Hope as Canada, Defence Research Board, T 176 R, April 1955)

Suppl.
Part. I
1956

LIST OF TABLES

Table

1. Mean monthly ABC types
2. Mean ABC types for entire period
3. Frequency of ABC types by month

3A. Cumulative ABC type frequencies

4. Frequency of ABC.D types for entire period
5. Frequency of observations for 23 months

Frequency distributions: comparisons of:

6–11: Surface elements

12–18: Surface elements by type and month

19–34: Inversion elements by type and month
35–56: Surface elements and inversion elements
57–62: Inversion elements

63. Mean annual lapse rates by type
64. Inversion lapse rate (R_1), and "intensity" (R_2) by surface temperature (T_1)
65. Extreme and usual values of variables for entire period
66. Percentage frequency of types by month (>10 per cent)

Note. Abbreviations are explained in Section IB of text.

INDEX TO FREQUENCY DISTRIBUTION TABLES

	ΔT			ΔP						
	a	*b*	*c*	*a*	*b*	*c*	*ABC*	*M*	Misc.	
T_1	36	37	38	39	40	41	12	16	P_1	6
W_1	42	43		44	45		13	17	T_1	7
									RH_1	8
									P_1	9
									P_1	46
P_1	47	48	49	50	51	52	14	18		
RH_1	53	54		55	56		15		P_1	11
									T_1	10
ΔTa		57		58	59		19	22		
b			60	61	62		20	23		
c							21	24		
ΔPa							25	28		
b							26	29		
c							27	30		
R_1							31	33	T_1	35
R_2							32	34		

Suppl.
Part. I
1956

Table 1. Mean Monthly ABC Types

Month	J	J	A	S	O	N	D	J	F	M	A	M
000 N	1	8	14	1								1
T_1	−2·0	+0·1	−2·5	−3·0								−7·0
P_1	993	995	994	1010								1005
001 N	4	5	11	12	1	1						2
T_1	−0·3	+0·2	−2·8	− 8·8	−22·0	−20·0						− 6·5
ΔT_c	−4·6	−4·4	−4·5	− 4·0	− 2·7	− 2·9						− 6·5
T_c	−4·9	−4·2	−7·3	−12·8	−24·7	−22·9						−13·0
P_1	1021	1002	1006	1009	1004	1012						1020
ΔP_c	231	190	213	203	292	293						272
P_c	790	812	793	806	712	719						748
002 N	1	3	4	1	2							
T_1	−2·0	+1·0	−0·7	−8·0	−15·0							
ΔT_c	−0·7	0·0	−0·5	−0·6	+ 0·1							
T_c	−2·7	+1·0	−1·2	−8·6	−14·9							
P_1	1022	1013	1009	1013	1014							
ΔP_c	266	82	181	87	189							
P_c	756	931	828	926	825							
030 N	2	6	2	2	2	17	4	9	6	7	4	
T_1	0·0	+1·1	−6·0	−13·5	−29·0	−30·0	−28·0	−33·0	−33·0	−43·0	−31·0	
ΔT_b	+2·4	+1·8	+5·4	+ 6·7	+11·4	+14·0	+ 9·4	+13·9	+14·6	+17·6	+16·5	
T_b	+2·4	+2·9	−0·6	− 6·8	−17·6	−16·0	−18·6	−19·1	−18·4	−25·4	−14·5	
P_1	1009	1000	1016	1016	1010	1012	1005	1019	1003	1013	1017	
ΔP_b	45	33	31	47	114	54	136	116	104	142	127	
P_b	964	967	985	969	896	958	869	903	899	871	890	

Table 1. Mean Monthly ABC Types (continued)

Month	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*
031 *N*	3	7	8	23	14	23	64	48	32	36	39	13
T_1	−0·3	+1·0	−7·6	−14·8	−27·0	−30·0	−33·0	−36·0	−38·0	−41·0	−27·0	−11·0
ΔT_b	+3·2	+3·4	+5·5	+ 8·0	+10·5	+11·6	+13·4	+14·0	+13·9	+16·3	+11·2	+ 2·3
T_b	+2·9	+4·4	−2·1	− 6·8	−16·5	−18·4	−19·6	−22·0	−24·1	−24·7	−15·8	− 8·7
ΔT_c	−3·8	−3·3	−3·4	− 3·2	− 2·9	− 3·0	− 3·4	− 2·7	− 3·4	− 2·7	− 4·0	− 6·2
T_c	−0·9	+1·1	−5·5	−10·0	−19·4	−21·4	−23·0	−24·7	−27·5	−27·4	−19·8	−14·9
P_1	1015	1006	1015	1012	1015	1015	1015	1014	1008	1017	1020	1019
ΔP_b	49	46	74	49	52	65	91	84	82	81	75	19
P_b	966	960	951	963	963	950	924	930	926	936	945	1000
ΔP_c	122	97	161	139	195	187	174	181	209	185	180	260
P_c	844	863	780	824	768	763	750	749	717	751	765	740
032 *N*		1	1	8	6	18	18	23	25	38	22	4
T_1		+2·0	−7·0	−17·1	−22·8	−32·0	−34·0	−37·0	−40·0	−40·0	−27·0	−12·0
ΔT_b		+3·7	+3·6	+ 7·5	+ 7·8	+12·5	+12·4	+13·8	+15·5	+15·0	+ 9·5	+ 3·8
T_b		+5·7	−3·4	− 9·6	−15·0	−19·5	−21·6	−23·2	−24·5	−25·0	−17·5	− 8·2
ΔT_c		−0·5	−0·9	− 0·1	− 0·3	+ 0·3	0·0	− 0·2	+ 0·1	0·0	+ 0·1	+ 0·2
T_c		+5·2	−4·3	− 9·7	−15·3	−19·2	−21·6	−23·4	−24·4	−25·0	−17·4	− 8·0
P_1		1001	1025	1011	1013	1013	1009	1012	1011	1013	1017	1016
ΔP_b		35	32	43	41	47	62	72	84	77	67	19
P_b		966	993	968	972	966	947	940	927	936	950	997
ΔP_c		54	137	77	74	125	141	149	145	120	110	88
P_c		912	856	891	898	841	806	791	782	816	840	909

Supp
Part
1956

Suppl.
Part. I
956

Table 1. Mean Monthly ABC Types (continued)

Month	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*
033 *N*				1	2	6	8	18	14	22	10	
T_1				−14·0	−30·0	−33·0	−31·0	−39·0	−39·0	−40·0	−27·0	
ΔT_b				+ 3·6	+ 8·0	+ 9·8	+11·5	+11·3	+10·5	+14·1	+10·5	
T_b				−10·4	−22·0	−23·2	−19·5	−27·7	−28·5	−25·9	−16·5	
ΔT_c				+ 2·3	+ 2·8	+ 2·7	+ 1·3	+ 2·2	+ 2·3	+ 1·8	+ 1·9	
T_c				− 8·1	−19·2	−20·5	−18·2	−25·5	−26·2	−24·1	−14·6	
P_1				1000	995	1009	1007	1014	1011	1016	1020	
ΔP_b				8	37	41	62	44	67	48	41	
P_b				992	958	968	945	970	944	968	979	
ΔP_c				189	220	147	70	170	163	122	105	
P_c				803	738	821	875	800	781	846	874	

101 *N*	2	12	10	4								2
T_1	+1·0	0·0	−1·5	− 4·8								− 6·5
ΔT_a	−3·1	−4·3	−4·7	− 1·4								− 5·6
T_a	−2·1	−4·3	−6·2	− 6·2								−12·1
ΔT_c	−3·6	−2·4	−2·3	− 4·5								− 1·6
T_c	−5·7	−6·7	−8·5	−10·7								−13·7
P_1	1018	1003	1005	1002								1012
ΔP_a	54	66	96	27								126
P_a	964	937	909	975								886
ΔP_c	200	140	132	196								118
P_c	764	797	777	779								768

Table 1. Mean Monthly ABC Types (continued)

Month	J	J	A	S	O	N	D	J	F	M	A	M
102 N	2	7	3	1							1	1
T_1	− 1·5	−0·4	−0·3	− 9·0							−14·0	−12·0
ΔT_a	− 8·3	−6·3	−3·2	− 0·9							− 1·8	− 5·5
T_a	− 9·8	−6·7	−3·5	− 9·9							−15·8	−17·5
ΔT_c	− 0·4	+0·3	+0·3	− 0·3							+ 0·5	+ 0·4
T_c	−10·2	−6·4	−3·2	−10·2							−15·3	−17·1
P_1	1004	1000	1003	1010							1016	1020
ΔP_a	108	114	57	9							31	132
P_a	896	886	946	1001							985	888
ΔP_c	71	83	59	107							97	128
P_c	825	803	887	894							888	760
130 N	21	17	8	10	3	7	2	2	1	1	4	5
T_1	−2·1	−0·1	−2·6	− 7·7	−17·7	−22·0	−28·0	−23·0	−23·0	−34·0	−16·0	− 7·6
ΔT_a	−3·9	−5·1	−3·4	− 2·3	− 2·1	− 5·2	− 0·6	− 2·8	− 5·2	− 2·2	− 2·7	− 6·1
T_a	−6·0	−5·2	−6·0	−10·0	−19·8	−27·2	−28·6	−25·8	−28·2	−36·2	−18·7	−13·7
ΔT_b	+3·5	+3·0	+1·7	+ 3·3	+ 7·8	+ 7·5	+14·3	+ 6·8	+17·7	+14·3	+ 5·5	+ 4·3
T_b	−2·5	−2·2	−4·3	− 6·7	−12·0	−19·7	−14·3	−19·0	−10·5	−21·9	−13·2	− 9·4
P	1006	1000	1005	1013	1012	1010	1015	1008	986	1008	1009	1014
ΔP_a	54	100	70	50	30	93	39	156	108	40	71	129
P_a	952	900	935	963	982	917	976	852	878	968	938	885
ΔP_b	39	37	31	54	67	83	144	93	64	217	54	33
P_b	913	863	904	909	915	834	832	759	814	751	884	852

Supp
Part
1956

uppl.
art. I
956

Table 1. Mean Monthly ABC Types (continued)

Month	J	J	A	S	O	N	D	J	F	M	A	M
131 N	23	35	18	20	9	6	3		7	2	13	13
T_1	−1·3	−0·1	−4·0	−10·1	−15·3	−25·0	−29·0		−32·0	−33·0	−19·0	− 6·2
ΔT_a	−4·5	−3·1	−4·1	− 2·1	− 2·9	− 3·1	− 1·9		− 1·0	− 0·6	− 2·5	− 3·5
T_a	−5·8	−3·2	−8·1	−12·2	−18·2	−28·1	−30·9		−33·0	−33·6	−21·5	− 9·7
ΔT_b	+3·0	+4·0	+3·4	+ 4·5	+ 4·7	+ 8·2	+10·8		+10·9	+11·0	+ 7·8	+ 3·9
T_b	−2·8	+0·8	−4·7	− 7·7	−13·5	−19·9	−20·1		−22·1	−22·6	−13·7	− 5·8
ΔT_c	−3·2	−2·9	−3·4	− 4·1	− 2·4	− 3·1	− 4·0		− 1·9	− 3·3	− 3·0	− 4·7
T_c	−6·0	−2·1	−8·1	−11·8	−15·9	−23·0	−24·1		−24·0	−25·9	−16·7	−10·5
P_1	1008	1005	1009	1007	1011	1010	1010		1000	1003	1015	1016
ΔP_a	67	47	55	33	59	55	33		45	38	62	65
P_a	941	958	954	974	952	955	977		955	965	953	951
ΔP_b	27	38	27	42	42	43	41		90	111	55	30
P_b	914	920	927	932	910	912	936		865	854	898	921
ΔP_c	130	117	161	159	117	134	270		127	179	124	156
P_c	784	803	766	773	793	778	666		738	675	774	765
132 N	14	7	6	4	6	7	8	3	3	1	8	9
T_1	−1·6	0·0	−3·8	−10·5	−14·3	−23·0	−28·0	−32·0	−30·0	−35·0	−19·0	− 7·0
ΔT_a	−3·6	−2·6	−4·2	− 2·0	− 7·4	− 3·4	− 1·4	− 0·8	− 2·3	− 2·8	− 2·6	− 6·0
T_a	−5·2	−2·6	−8·0	−12·5	−21·7	−26·4	−29·4	−32·8	−32·3	−37·8	−21·6	−13·0
ΔT_b	+3·8	+4·2	+4·0	+ 3·8	+ 7·9	+ 5·3	+ 8·0	+ 8·5	+14·0	+10·1	+ 8·9	+ 3·5
T_b	−1·4	+1·6	−4·0	− 8·7	−13·8	−21·1	−21·4	−24·3	−18·3	−27·7	−12·7	− 9·5
ΔT_c	0·0	−0·1	0·0	+ 0·1	+ 0·6	0·0	0·0	− 0·1	− 0·2	+ 0·2	− 0·2	+ 0·1
T_c	−1·4	+1·5	−4·0	− 8·6	−13·2	−21·1	−21·4	−24·4	−18·5	−27·5	−12·9	− 9·4
P_1	1007	1008	1010	1003	1016	1010	1005	1005	994	1017	1007	1017
ΔP_a	42	35	57	62	125	62	31	17	65	34	57	122
P_a	965	973	953	941	891	948	974	988	929	983	950	895
ΔP_b	29	35	17	38	41	46	35	61	61	55	47	18
P_b	936	938	936	903	850	902	939	927	868	928	903	877
ΔP_c	62	68	84	52	48	81	123	134	76	148	76	79
P_c	874	870	852	851	802	821	816	793	792	780	827	798

Table 1. Mean Monthly ABC Types (continued)

Month	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*
133 *N*	1			2	3	4	2			1	2	3
T_1	0·0			− 4·5	−24·7	−25·0	−30·0			−33·0	−27·0	−13·0
ΔT_a	−0·8			− 6·7	− 1·9	− 1·9	− 2·5			− 0·7	− 2·6	− 2·2
T_a	−0·8			−11·2	−26·6	−26·9	−32·5			−33·7	−29·6	−15·2
ΔT_b	+3·5			+ 2·7	+ 5·3	+ 8·0	+ 8·5			+ 4·1	+ 7·4	+ 3·4
T_b	+2·7			− 8·5	−21·3	−18·9	−24·0			−29·6	−22·2	−11·8
ΔT_c	+1·0			+ 1·3	+ 2·8	+ 1·9	+ 2·6			+ 2·4	+ 2·2	+ 2·9
T_c	+3·7			− 7·2	−18·5	−17·0	−21·4			−27·2	−20·0	− 8·9
P_1	1019			1003	1000	1008	998			996	1006	1023
ΔP_a	29			135	25	34	58			68	56	46
P_a	990			868	975	974	940			928	950	977
ΔP_b	40			11	25	47	37			29	17	13
P_b	950			857	950	927	903			899	933	964
ΔP_c	72			83	138	99	112			105	152	142
P_c	878			774	812	828	791			794	781	822

Sup
Par
195

Suppl.
Part. I
1956

Table 2. Mean ABC Types for Entire Period

ABC	*000*	*001*	*002*	*030*	*031*	*032*	*033*	*101*	*102*	*130*	*131*	*132*	*133*
N	25	36	11	61	310	164	80	30	15	81	149	76	18
T_1	−2	−5	−4	−27	−30	−33	−36	−2	−3	−8	−9	−13	−20
ΔT_a								−4	−5	−4	−3	−4	−2
T_a								−6	−8	−12	−12	−17	−22
ΔT_b				+12	+12	+13	+12			+5	+5	+6	+6
T_b				−15	−18	−20	−24			−7	−7	−11	−16
ΔT_c		−4	0		−3	0	+2	−3	0		−3	0	+2
T_c		−9	−4		−21	−20	−22	−9	−8		−10	−11	−14
P_1	995	1009	1012	1011	1014	1012	1013	1005	1004	1007	1009	1009	1007
ΔP_a								−74	−91	−75	−52	−62	−52
P_a								931	913	932	957	947	955
ΔP_b				−92	−75	−66	−49			−51	−40	−35	−28
P_b				919	939	946	964			881	917	912	927
ΔP_c		−216	−138		−181	−125	−136	−147	−82		−139	−79	−117
P_c		793	874		758	821	828	784	831		778	833	810

Table 3. Frequency of ABC Types by Month

Month	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	% >1%	*N*
ABC														
000	1	8	14	1								1	2	25
001	4	5	11	12	1	1						2	3	36
002	1	3	4	1	2								1	11
003											1			1
030	2	6	2	2	2	17	4	9	6	7	4		6	61
031	3	7	7	23	14	23	64	48	33	36	39	13	28	310
032		1	1	8	6	18	18	23	25	38	22	4	15	164
033				1	2	6	8	18	13	22	10		7	80
100	1	2												3
101	2	12	10	4								2	3	30
102	2	7	3	1							1	1	1	15
103	2		1	1		2						3		9
130	21	17	8	10	3	7	2	2	1	1	4	5	7	81
131	23	35	18	20	9	6	3		7	2	13	13	14	149
132	14	7	6	4	6	7	8	3	3	1	8	9	7	76
133	1			2	3	4	2			1	2	3	2	18
201			1											1
230				3			1			1				5
231					1	2		3	2		2			10
232				1	1		1	2	1		1			7
233							1	2	1		1			5
N	77	110	86	94	50	93	112	110	92	109	108	56	100	1097

Table 3A. Cumulative ABC Type Frequencies

ABC types	%	*Cum.* %	*ABC types*	%	*Cum.* %
031	28	28	001	3	87
032	15	43	101	3	90
131	14	57	000	2	92
033	7	64	133	2	94
130	7	71	102	1	95
132	7	78	002	1	96
030	6	84	8 Misc.	4	100

Table 4. Frequency of ABC.D Types for Entire Period

D	*0*	*1*	*2*	*3*	*4*	*5*	*6*	*7*	*N*
ABC									
000	25								25
001	18			18					36
002	5			6					11
003				1					1
030	37		24						61
031	138		68	80		24			310
032	96		40	24		4			164
033	37		20	18		5			80
100		3							3
101	20			7			3		30
102	10	1		3			1		15
103	5	1		3					9
130	47	13	18		3				81
131	86	16	13	28	2	1	2	1	149
132	43	9	8	12		1	2	1	76
133	6	2		8			2		18
201	1								1
230	1		4						5
231	4	1		3	2				10
232	2	3	1			1			7
233	2			1		1		1	5
N	583	49	196	212	7	37	10	3	1097

Suppl.
Part. I
1956

Table 5. Observations at T-3, 21 June 1952–26 April 1954

	1952	*1953*	*1954*	*Total*		*1952*	*1953*	*1954*	*Total*
January		55	55	110	July	50	60		110
February		54	38	92	August	31	55		86
March		54	55	109	September	42	52		94
April		59	49	108	October	11	39		50
May		56		56	November	42	51		93
June	19	58		77	December	59	53		112
									1097

Table 6. T_1—P_1

P_1 (*mb*)	*970 979*	*980 989*	*990 999*	*1000 1009*	*1010 1019*	*1020 1029*	*1030 1039*	*1040 1049*	*N*
T_1 (°C)									
+2 +1		2	12	16	22	2			54
0	2	2	12	34	26	4			80
−1 −02		3	16	27	17	8			71
−3 −09		5	24	42	52	19			142
−10 −14			3	20	30	24	1		78
−15 −19			2	20	14	11			47
−20 −24		2	9	27	23	14			75
−25 −29		1	13	30	40	27	6		117
−30 −34		4	17	48	47	19	5		140
−35 −39		2	4	46	68	24	11	1	156
−40 −51			2	34	66	27	8		137
N	2	21	114	344	405	179	31	1	1097

Table 7. W_1—T_1

T_1 (°C)	+2 +1	0	−1 −2	−3 −9	−10 −14	−15 −19	−20 −24	−25 −29	−30 −34	−35 −39	−40 −44	−45 −51	N
W_1 (mps)													
00	6	5	6	13	11	9	6	8	9	12	10	5	100
01 02	2	6	5	8	8	9	9	18	27	43	25	16	176
03 05	38	43	31	50	33	14	39	61	66	85	60	15	535
06 08	6	19	20	49	11	12	13	21	26	15	6		198
09 11	1	7	8	19	13	3	6	7	7	1	1		73
12 14			1	1			1	1	3				7
15 16					2				1				3
N	53	80	71	140	78	47	74	116	139	156	102	36	1092

Table 8. W_1—RH_1

RH_1 (%)	20 29	30 39	40 49	50 59	60 69	70 79	80 89	90 99	N
W_1 (mps)									
00		3	9	15	19	17	14	2	79
01 02		6	9	25	36	25	13	5	119
03 05	1	6	15	47	84	129	103	57	442
06 08			2	7	24	64	53	39	189
09 11					4	19	30	18	71
12 14				1	1	3		2	7
15 16						1	1	1	3
N	1	15	35	95	168	258	214	124	910

Table 9. W_1—P_1

P_1 (mb)	970 979	980 989	990 999	1000 1009	1010 1019	1020 1029	1030 1039	1040 1049	N
W_1 (mps)									
00		3	6	34	37	16	4		100
01 02				54	79	36	7		176
03 05	1	8	59	162	195	91	19		535
06 08		7	32	69	60	29		1	198
09 11	1	2	15	20	29	5	1		73
12 14			2	2	2	1			7
15 16				3					3
N	2	20	114	344	402	178	31	1	1092

Table 10. RH_1—T_1

T_1 (°C)	+2 +1	0	−1 −2	−3 −9	−10 −14	−15 −19	−20 −24	−25 −29	−30 −34	−35 −39	−40 −44	N
RH (%)												
20 29										1		1
30 39								2	2	10	1	15
40 49						1	2	6	12	15		36
50 59				2	9	7	5	23	26	23	1	96
60 69	1	2		7	12	12	24	29	47	28	6	168
70 79	14	9	6	27	32	13	23	55	50	30	1	260
80 89	22	30	35	70	24	13	21					215
90 99	17	39	30	36	1			1				124
N	54	80	71	142	78	46	75	116	137	107	9	915

Table 11. RH_1—P_1

P_1 (*mb*)	970 979	980 989	990 999	1000 1009	1010 1019	1020 1029	1030 1039	1040 1049	N
RH_1 (%)									
20 29				1					1
30 39				6	7	2			15
40 49			2	13	15	6			36
50 59			7	34	32	21	2		96
60 69		1	22	51	61	29	3	1	168
70 79		11	25	78	87	49	10		260
80 89		5	28	78	73	31			215
90 99	2	4	28	39	44	7			124
N	2	21	112	300	319	145	15	1	915

Table 12. T_1—ABC

T_1 (°C)	+2 +1	0	−1 −2	−3 −9	−10 −14	−15 −19	−20 −24	−25 −29	−30 −34	−35 −39	−40 −44	−45 −51	N
ABC													
000	2	6	8	9									25
001	2	7	6	15	3	1	2						36
002	3	3	1	2	1	1							11
003							1						1
030	6	2	1	1	1	2	6	11	10	13	5	3	61
031	7	2		13	23	14	16	50	58	75	39	13	310
032	1			4	4	6	9	22	34	33	38	13	164
033					2	1	2	8	16	26	19	6	80
100	2	1											3
101	5	11	4	10									30
102	2	3	7	1	2								15
103			1	6	1	1							9
130	7	14	14	22	7	5	5	6	1				81
131	13	24	20	41	16	9	12	4	8	1	1		149
132	4	6	9	14	12	4	12	8	3	4			76
133		1		2	3	1	4	3	4				18
201				1									1
230				1	2		1			1			5
231					1		4	1	4				10
232						2		2	2	1			7
233							1	2		2			5
N	54	80	71	142	78	47	75	117	140	156	102	35	1097

Suppl. Part. I 1956

Table 13. W_1—ABC

W_1 (*mps*)	*00*	*01* *02*	*03* *05*	*06* *08*	*09* *11*	*12* *14*	*15* *16*	N
ABC								
000	4		12	8	1			25
001	6	4	15	8	3			36
002			8	1	1			10
003				1				1
030	7	11	39	1	3			61
031	26	74	163	30	10	3	1	307
032	16	38	86	20	3	1		164
033	7	21	40	12				80
100			2	1				3
101	1	2	16	5	6			30
102		1	8	5	1			15
103	1		3	4	1			9
130	8	7	35	21	9	1		81
131	13	10	64	45	15		1	148
132	6	5	22	24	17	1	1	76
133	3	3	7	3	2			18
201			1					1
230			3	2				5
231	1		5	2	1	1		10
232			3	4				7
233	1		3	1				5
N	100	176	535	198	73	7	3	1092

Table 14. P_1—ABC

P_1 (*mb*)	*970* *979*	*980* *989*	*990* *999*	*1000* *1009*	*1010* *1019*	*1020* *1029*	*1030* *1039*	*1040* *1049*	N
ABC									
000	1	6	9	7	2				25
001			6	10	15	5			36
002				2	8	1			11
003					1				1
030			7	22	19	11	1	1	61
031		3	19	62	127	79	20		310
032		2	12	45	73	28	4		164
033			3	24	35	12	6		80
100			1		2				3
101	1		7	12	7	3			30
102		1	2	7	4	1			15
103			3	4	2				9
130		3	11	36	25	6			81
131		5	20	60	45	19			149
132		1	6	34	29	6			76
133			5	6	4	3			18
201			1						1
230			1	2		2			5
231				5	3	2			10
232				3	3	1			7
233			1	3	1				4
N	2	21	114	344	405	179	31	1	1097

Table 15. RH_1—ABC

RH_1 (%)	*20* *29*	*30* *39*	*40* *49*	*50* *59*	*60* *69*	*70* *79*	*80* *89*	*90* *99*	*N*
ABC									
000					4	6	9	9	28
001				2		4	18	9	33
002					2	1	2	6	11
003						1			1
030		1	4	4	9	21	6	2	47
031	1	7	12	45	66	83	25	3	242
032		5	12	22	26	32	5		102
033		2	2	11	18	13			46
100							1	2	3
101						3	12	15	30
102						3	9	3	15
103					1	4	3	1	9
130				1	6	18	37	18	80
131			3	2	11	36	59	38	149
132				1	12	25	20	17	75
133			1	4	4	4	4	1	18
201							1		1
230					1	2	1		4
231			1	3	2	2	2		10
232					5	1	1		7
233			1	1	1	1			4
N	1	15	36	96	168	260	215	124	915

Table 16. T_1—M

Month		*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	*N*
T_1	(°C)													
+2	+1	13	40	1										54
0		16	46	17									1	80
−1	−2	27	22	18	2								2	71
−3	−9	21	2	47	40								32	142
−10	−14			3	33	13	1					11	17	78
−15	−19				12	8	6	1	1			15	4	47
−20	−24				6	9	16	9	7	5		23		75
−25	−29				1	13	30	27	11	4	1	30		117
−30	−34					4	22	31	25	26	15	17		140
−35	−39					3	18	37	40	21	28	9		156
−40	−44							7	17	25	50	3		102
−45	−51								9	11	15			35
N		77	110	86	94	50	93	112	110	92	109	108	56	1097

Table 17. W_1—M

Month	J	J	A	S	O	N	D	J	F	M	A	M	N
W_1 (mps)													
00	5	9	12	12	7	13	8	11	3	5	8	7	100
01 02		9	6	11	8	22	32	18	20	26	15	9	176
03 05	43	57	40	37	23	41	51	58	51	63	55	16	535
06 08	21	23	25	19	9	11	13	17	15	11	21	13	198
09 11	7	11	3	12	3	5	7	4	1	3	7	10	73
12 14				1		1	1	2		1		1	7
15 16				1					1		1		3
N	76	109	86	93	50	93	112	110	91	109	107	56	1092

Table 18. P_1—M

Month	J	J	A	S	O	N	D	J	F	M	A	M	N
P_1 (mb)													
970 979		2											2
980 989	2	6	4		1			1	7				21
990 999	12	27	19	8	3	11	11	5	13	5			114
1000 1009	23	48	29	39	12	30	37	35	31	30	22	8	344
1010 1019	30	27	27	32	24	28	43	44	29	44	47	30	405
1020 1029	10		7	15	10	21	19	18	12	17	33	17	179
1030 1039						3	2	6		13	6	1	31
1040 1049								1					1
N	77	110	86	94	50	93	112	110	92	109	108	56	1097

Table 19. ΔT_a—ABC

ΔT_a (°C)	+2·0 +0·1	0·0 −1·9	−2·0 −3·9	−4·0 −5·9	−6·0 −7·9	−8·0 −9·9	−10·0 −11·9	−12·0 −13·9	−14·0 −15·9	−16·0 −16·6	N
ABC											
100						1	1	1			3
101		8	9	6	3	3	1				30
102		2	5	3	1	2	2				15
103		2	2	3	2						9
130		24	23	21	7	1	2	2		1	81
131		54	52	24	11	7	1				149
132	2	24	29	9	3	3	5	1			76
133		9	7		2						18
201		1									1
230	1	4									5
231	4	6									10
232	4	3									7
233	2	3									5
N	13	140	127	66	29	17	12	4		1	409

Table 20. ΔT_b—*ABC*

ΔT_b (+°C)	*30·5* *30·0*	*29·9* *26·0*	*25·9* *22·0*	*21·9* *18·0*	*17·9* *14·0*	*13·9* *10·0*	*9·9* *6·0*	*5·9* *2·0*	*1·9* *0·0*	*N*
ABC										
030	1		2	7	18	10	10	9	4	61
031		3	7	37	66	86	60	40	11	310
032			3	21	39	53	35	11	2	164
033				9	17	23	19	11	1	80
130					3	3	15	34	26	81
131					4	8	30	87	20	149
132					2	8	23	38	5	76
133						2	5	11		18
230						1	2	2		5
231						4	5	1		10
232				1		1	3	2		7
233						1	2	2		5
N	1	3	12	75	149	200	209	248	69	966

Table 21. ΔT_c—*ABC*

ΔT_c (°C)	*+5·9* *+2·0*	*+1·9* *−1·9*	*−2·0* *−5·9*	*−6·0* *−9·9*	*−10·0* *−12·4*	*N*
ABC						
001		9	17	9	1	36
002		11				11
003	1					1
031	1	92	179	34	4	310
032		164				164
033	36	44				80
101		14	14	2		30
102		15				15
103	2	7				9
131		46	81	20	2	149
132		76				76
133	10	8				18
201				1		1
231		5	5			10
232		7				7
233	2	3				5
N	52	501	296	66	7	922

Table 22. ΔT_a—*M*

Month		*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	*N*
ΔT_1	(°C)													
+2·0	+0·1				2			1	5	3		2		13
0·0	−1·9	14	21	7	27	10	7	14	5	9	4	13	9	140
−2·0	−3·9	21	31	19	10	7	12	3	2	1	2	13	6	127
−4·0	−5·9	18	10	13	4	2	4			2		3	10	66
−6·0	−7·9	4	7	3	3	2	1					1	8	29
−8·0	−9·9	4	4	4		2	1						2	17
−10·0	−11·9	2	6	1		2	1							12
−12·0	−13·9	3											1	4
−14·0	−15·9													
−16·0	−16·6		1											1
N		66	80	47	46	25	26	18	12	15	6	32	36	409

Table 23. ΔT_b—*M*

Month	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	*N*
ΔT_b (+°C)													
30·5 30·0											1		1
29·9 26·0									2		1		3
25·9 22·0							1		3	6	2		12
21·9 18·0					2	6	11	12	14	28	2		75
17·9 14·0				2	2	22	24	34	22	38	5		149
13·9 10·0		1		8	10	19	40	36	28	20	38		200
9·9 6·0	8	9	8	19	14	24	30	22	16	11	44	4	209
5·9 2·0	38	42	24	38	18	19	6	5	7	5	11	35	248
1·9 0·0	18	21	10	7	1			1		1	2	8	69
N	64	73	42	74	47	90	112	110	92	109	106	47	966

Table 24. ΔT_c—*M*

Month	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	*N*
ΔT_c (°C)													
+5·9 +2·0				1	5	7	3	9	8	12	5	2	52
+1·9 −1·9	29	42	27	36	23	44	54	60	50	59	54	23	501
−2·0 −5·9	18	28	28	28	16	15	42	26	23	28	31	13	296
−6·0 −9·9	4	7	6	12	1	2	6	4	2	1	10	11	66
−10·0 −12·4	1		1	1		1			2			1	7
N	52	77	62	78	45	69	105	99	85	100	100	50	922

Table 25. ΔP_a—*ABC*

ΔP_a (−*mb*)	*00 19*	*20 39*	*40 59*	*60 79*	*80 99*	*100 119*	*120 139*	*140 159*	*160 179*	*180 199*	*200 292*	*N*
ABC												
100											3	3
101	4	9	1	3	3	4	1	3		2		30
102	1	1	4	2	1	1	2	2			1	15
103		3		2	2	1	1					9
130	11	17	11	13	6	8	4	2	1	4	4	81
131	21	46	30	28	9	6	4	3	1		1	149
132	14	27	9	5	4	6	4	2	1	3	1	76
133	1	8	4	1	3				1			18
201			1									1
230	2	2		1								5
231	1	5	1	1				1			1	10
232	1	2	2	1	1							7
233	1	2	1	1								5
N	57	122	64	58	29	26	16	13	4	9	11	409

Table 26. ΔP_b—*ABC*

ΔP_b (−*mb*)	*00* *19*	*20* *39*	*40* *59*	*60* *79*	*80* *99*	*100* *119*	*120* *139*	*140* *159*	*160* *179*	*180* *199*	*200* *246*	N
ABC												
030	4	8	15	4	6	4	5	4	6	2	3	61
031	29	49	50	44	50	44	25	6	6	2	5	310
032	17	30	27	33	23	22	6	3	1		2	164
033	12	22	17	17	10	1	1					80
130	15	26	18	6	5	3	3	3		1	1	81
131	42	51	22	20	8	1	2	1	2			149
132	27	23	13	11	2							76
133	7	9	1			1						18
230			2	1		1					1	5
231	1	3	1	5								10
232	1	2	1	1	1	1						7
233		1	3		1							5
N	155	224	170	142	106	78	42	17	15	5	12	966

Table 27. ΔP_c—*ABC*

ΔP_c (−*mb*)	*00* *99*	*100* *199*	*200* *299*	*300* *399*	*400* *437*	N
ABC						
001	1	14	17	4		36
002	4	5	2			11
003			1			1
031	39	150	102	18	1	310
032	75	59	28	2		164
033	31	36	12	1		80
101	3	21	6			30
102	10	5				15
103	5	2	2			9
131	50	67	30	2		149
132	55	19	1	1		76
133	7	10	1			18
201		1				1
231	3	4	2	1		10
232	4	3				7
233	1	3	1			5
N	288	399	205	29	1	922

Table 28. ΔP_a—*M*

Month	*J*	*F*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	N
ΔP_a (−*mb*)													
00 19	11	14	3	15	1	3	3	3	1		3		57
20 39	17	17	10	18	9	12	9	3	5	3	13	6	122
40 59	11	13	11	2	4	3	4	1	4	2	4	5	64
60 79	11	15	11	2	2	4	1	2	2	1	3	4	58
80 99	6	2	4	5		3	1				4	4	29
100 119	3	6	3	2				1	2		2	7	26
120 139	2	2		1	4						1	6	16
140 159	1	4	1		2	1			1		2	1	13
160 179			2	1								1	4
180 199	1	2	2		1	2						1	9
200 292	3	5						2				1	11
N	66	80	47	46	23	28	18	12	15	6	32	36	409

Suppl.
Part. I
1956

Table 29. ΔP_b—*M*

Month		*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	*N*
ΔP_b (−*mb*)														
00	19	18	23	12	18	12	11	6	7	4	9	12	23	155
20	39	25	22	20	22	13	20	18	18	9	15	26	16	224
40	59	15	11	6	14	10	21	13	16	26	18	15	5	170
60	79	4	11	1	11	5	16	19	17	16	20	19	3	142
80	99	2	5	1	5	3	9	18	21	13	17	12		106
100	119		1	1			6	22	14	7	13	14		78
120	139				2	3	5	9	11	6	5	1		42
140	159			1	1			3	4	3	1	4		17
160	179					1	2	1	2	2	6	1		15
180	199				1			2		1		1		5
200	246							1		5	5	1		12
N		64	73	42	74	47	90	112	110	92	109	106	47	966

Table 30. ΔP_c—*M*

Month		*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	*N*
ΔP_c (−*mb*)														
00	99	25	32	18	22	18	25	29	23	23	29	34	10	288
100	199	16	37	31	39	15	24	48	43	32	47	46	21	399
200	299	10	8	12	15	12	17	23	28	22	23	19	16	205
300	399	1		1	2		3	5	4	8	1	1	3	29
400	437								1					1
N		52	77	62	78	45	69	105	99	85	100	100	50	922

Table 32. R_2—*ABC*

$R_2 = \frac{(\Delta T_b)^2}{\Delta P_b}$	*0·0* *0·3*	*0·4* *1·1*	*1·2* *1·9*	*2·0* *3·9*	*4·0* *5·9*	*6·0* *9·9*	*10·0* *13·9*	*14·0* *17·9*	*18·0* *33·0*	*N*
ABC										
030	12	14	9	15	6	5	1			62
031	32	63	58	97	37	17	2	2	3	311
032	9	19	31	52	27	21	2	1	2	164
033	3	9	15	26	9	11	5	1		79
130	45	23	7	2	2					79
131	58	53	22	15		1				149
132	14	35	11	15	1	2				78
133	2	9	2	3	2					18
230	3	1	1							5
231	1	2	3	3						9
232	2	1	1	1	1		1			7
233	1	2	1	1						5
N	182	231	161	230	85	57	11	4	5	966

Suppl.
Part. I
1956

Table 31. R_1—ABC

$R_1 = \frac{\Delta T_b}{\Delta P_b}$	*0·00* *0·04*	*0·05* *0·09*	*0·10* *0·14*	*0·15* *0·19*	*0·20* *0·24*	*0·25* *0·29*	*0·30* *0·39*	*0·40* *0·49*	*0·50* *0·59*	*0·60* *0·69*	*0·70* *0·79*	*0·80* *0·89*	*0·90* *0·99*	*1·00* *1·49*	*1·50* *2·01*	*N*
ABC																
030	10	11	13	7	5	2	10	3		1						62
031	12	44	83	61	42	16	19	16	5	4	4	1		3	1	311
032	1	17	27	36	25	14	16	9	6	4	3	1	1	2	2	164
033	3	1	4	23	12	6	10	9	6	1	2	1	1			79
130	20	27	14	8	3	3	2		1			1				79
131	14	35	44	16	14	5	14	5	2							149
132	2	10	17	15	14	7	5	5	1	1	1					78
133		2	2	4	3	1	3	1	2							18
230	1	3		1												5
231			5	1		1	2									9
232	2			1	1	1	1							1		7
233		1	2	1	1											5
N	65	151	211	174	120	56	82	48	23	11	10	4	2	6	3	966

Table 33. R_1—M

Month	J	J	A	S	O	N	D	J	F	M	A	M	N
$R_1 = \frac{\Delta T_b}{\Delta P_b}$													
·00 ·04	13	16	7	6	1	3	10	4	2		1	2	65
·05 ·09	20	22	12	17	8	14	15	9	9	3	16	6	151
·10 ·14	11	19	8	15	6	11	26	28	16	24	33	14	211
·15 ·19	7	4	6	14	8	15	20	25	24	23	17	11	174
·20 ·24	7	3	1	6	9	12	16	15	17	18	12	4	120
·25 ·29	2	3	4	4	2	9	6	4	8	9	5		56
·30 ·39	2	4	4	6	4	11	8	7	11	14	7	5	83
·40 ·49	1	1		4	5	6	1	9	2	4	9	5	47
·50 ·59		1		1	1	3	4	3	1	6	3		23
·60 ·69					3	2	2	1			3		11
·70 ·79						1	2	1		6			10
·80 ·89	1					1	1	1		·			4
·90 ·99								2					2
1·00 1·49				1		1	1	1	2	1			7
1·50 2·01						1				1			2
N	64	73	42	74	47	90	112	110	92	109	106	47	966

Table 34. R_2—M

Month	J	J	A	S	O	N	D	J	F	M	A	M	N
$R_2 = \frac{(\Delta T_b)^2}{\Delta P_b}$													
0·0 0·3	41	41	22	24	7	8	7	5	2	1	5	19	182
0·4 1·1	14	24	13	24	14	23	27	15	11	10	35	21	231
1·2 1·9	7	5	3	15	8	14	22	22	20	16	26	3	161
2·0 3·9	1	3	4	8	10	23	32	39	40	42	25	4	231
4·0 5·9	1				6	7	14	14	15	19	8		84
6·0 9·9				2	2	11	9	11	2	13	7		57
10·0 13·9						1		2	1	6			10
14·0 17·9				1		1		2	1				5
18·0 33·0						2	1			2			5
N	64	73	42	74	47	90	112	110	92	109	106	47	966

Table 35. R_1—T_1

T_1 (°C)		+2 +1	0 −4	−5 −9	−10 −14	−15 −19	−20 −24	−25 −29	−30 −34	−35 −39	−40 −44	−45 −49	−50 −51	N
$R_1 = \frac{\Delta T_b}{\Delta P_b}$														
·00	·04	5	25	5	2	1	4	5	1	3				51
·05	·09	15	37	15	13	8	14	25	17	10	6	1		161
·10	·14	7	29	15	18	7	18	27	35	31	15	8		210
·15	·19	3	13	9	13	8	13	15	30	33	23	9	1	170
·20	·24	3	10	4	9	5	8	16	19	26	19	7	1	127
·25	·29	1	6	4	6	4	4	9	6	10	11	3		64
·30	·39	3	5	8	3	4	5	9	13	14	11	1		76
·40	·49	1	1	3	6	5	1	4	8	13	3	2		47
·50	·59		1		1	1	3	1	5	5	5	1		23
·60	·69						2	2	3	3	1			11
·70	·79							2	2	2	4			10
·80	·89			1				1		1	1			4
·90	·99							1			1	1		3
1·00	1·49					1				4	1			6
1·50	2·01								1	1	1			3
N		38	127	64	71	44	72	117	140	156	102	33	2	966

Table 36. T_1—ΔT_a

ΔT_a (°C)		+2·0 +0·1	0·0 −1·9	−2·0 −3·9	−4·0 −5·9	−6·0 −7·9	−8·0 −9·9	−10·0 −11·9	−12·0 −13·9	−14·0 −15·9	−16·0 −16·6	N
T_1	(°C)											
+2	+1		11	8	8	2	2	1	1			33
0			17	20	7	6	5	2	2		1	60
−1	−02		2	26	17	4	1	5				55
−3	−09	1	31	27	20	11	6	1	1			98
−10	−14		22	8	6	5	1	2				44
−15	−19	2	4	11	2		2	1				22
−20	−24	4	18	12	5							39
−25	−29		15	9	1	1						26
−30	−34	3	14	5								22
−35	−39	3	5	1								9
−40	−44		1									1
N		13	140	127	66	29	17	12	4		1	409

Table 37. T_1—ΔT_b

ΔT_b (+°C)	*30·5 30·0*	*29·9 26·0*	*25·9 22·0*	*21·9 18·0*	*17·9 14·0*	*13·9 10·0*	*9·9 6·0*	*5·9 2·0*	*1·9 0·0*	N
T_1 (°C)										
+2 +1						1	1	26	10	38
0							5	29	15	49
–1 –02							4	27	13	44
–3 –09							18	60	20	98
–10 –14						7	22	36	6	71
–15 –19					2	4	14	21	3	44
–20 –24					4	20	31	17		72
–25 –29				1	11	49	40	15	1	117
–30 –34			1	10	33	41	44	11		140
–35 –39	1	1	3	28	48	49	19	6	1	156
–40 –51		2	8	36	51	29	11			137
N	1	3	12	75	149	200	209	248	69	966

Table 38. T_1—ΔT_c

ΔT_c (°C)	*+5·9 +2·0*	*+1·9 –1·9*	*–2·0 –5·9*	*–6·0 –9·9*	*–10·0 –12·4*	N
T_1 (°C)						
+2 +1		20	16	1		37
0		31	18	8		57
–1 –02		29	16	3		48
–3 –09		48	43	15	3	109
–10 –14	4	32	23	9		68
–15 –19	1	22	9	8		40
–20 –24	3	33	24	3		63
–25 –29	8	53	33	5	1	100
–30 –34	15	70	34	9	1	129
–35 –39	8	81	49	3	1	142
–40 –51	13	82	31	2	1	129
N	52	501	296	66	7	922

Table 39. T_1—ΔP_a

ΔP_a (–*mb*)	*00 19*	*20 39*	*40 59*	*60 79*	*80 99*	*100 119*	*120 139*	*140 159*	*160 179*	*180 199*	*200 292*	N
T_1 (°C)												
+2 +1	7	7	4	6	2	3	1	1			2	33
0	6	13	9	14	4	3	2	1		4	4	60
–1 –02	8	13	8	10	2	4	4	3		1	2	55
–3 –09	16	23	17	12	12	8	2	2	4	1	1	98
–10 –14	5	15	4	6	2	3	7	1		1		44
–15 –19	1	10	2	1	1	3		2		2		22
–20 –24	6	18	6	4	3	1		1				39
–25 –29	3	9	8	1	2	1		1			1	26
–30 –34	1	9	5	4	1			1			1	22
–35 –39	4	5										9
–40 –44			1									1
N	57	122	64	58	29	26	16	13	4	9	11	409

Table 40. T_1—ΔP_b

ΔP_b (−*mb*)	*00* *19*	*20* *39*	*40* *59*	*60* *79*	*80* *99*	*100* *119*	*120* *139*	*140* *159*	*160* *179*	*180* *199*	*200* *246*	N
T_1 (°C)												
+2 +1	12	18	5	3								38
0	12	15	12	5	4	1						49
−1 −02	14	17	5	6	2							44
−3 −09	37	30	18	7	3	1		2				98
−10 −14	20	23	12	12	2		1			1		71
−15 −19	13	17	6	1	5	2						44
−20 −24	8	11	16	17	7	9	4					72
−25 −29	9	28	20	13	15	15	10	3	3	1		117
−30 −34	10	29	21	29	20	14	7	4	4		2	140
−35 −39	12	21	33	28	26	16	6	2	3	3	6	156
−40 −51	8	15	22	21	22	20	14	6	5		4	137
N	155	224	170	142	106	78	42	17	15	5	12	966

Table 41. T_1—ΔP_c

ΔP_c (−*mb*)	*00* *99*	*100* *199*	*200* *299*	*300* *437*	N
T_1 (°C)					
+2 +1	18	16	3		37
0	19	28	10		57
−1 −02	22	19	6	1	48
−3 −09	32	49	26	2	109
−10 −14	22	31	13	2	68
−15 −19	14	15	9	2	40
−20 −24	31	20	12		63
−25 −29	31	47	20	2	100
−30 −34	37	51	34	7	129
−35 −39	35	60	38	9	142
−40 −51	27	63	34	5	129
N	288	399	205	30	922

Table 42. W_1—ΔT_a

ΔT_a (°C)	+*2·0* +*0·1*	*0·0* −*1·9*	−*2·0* −*3·9*	−*4·0* −*5·9*	−*6·0* −*7·9*	−*8·0* −*9·9*	−*10·0* −*11·9*	−*12·0* −*13·9*	−*14·0* −*15·9*	−*16·0* −*16·6*	N
W_1 (mps)											
00	1	6	11	6	5	1	3	1			34
01 02		10	5	8	3	1				1	28
03 05	6	68	51	22	10	7	6	2			172
06 08	5	37	39	18	9	7	2				117
09 11		18	19	11	2		1	1			52
12 14	1		1	1							3
15 16		1	1								2
N	13	140	127	66	29	16	12	4		1	408

Table 43. W_1—ΔT_b

ΔT_b (+°C)	*30·5* *30·0*	*29·9* *26·0*	*25·9* *22·0*	*21·9* *18·0*	*17·9* *14·0*	*13·9* *10·0*	*9·9* *6·0*	*5·9* *2·0*	*1·9* *0·0*	N
W_1 (mps)										
00			2	7	10	16	19	31	3	88
01 02		2	6	19	37	42	30	32	1	169
03 05	1	1	4	46	75	105	102	100	36	470
06 08				2	22	29	33	58	21	165
09 11					4	6	20	25	5	60
12 14					1	1	2		3	7
15 16				1			1	1		3
N	1	3	12	75	149	199	207	247	69	962

Table 44. W_1—ΔP_a

ΔP_a (−*mb*)	*00* *19*	*20* *39*	*40* *59*	*60* *79*	*80* *99*	*100* *119*	*120* *139*	*140* *159*	*160* *179*	*180* *199*	*200* *292*	N
W_1 (mps)												
00	2	11	1	5	2	3	4		1	3	2	34
01 02	1	6	7	4	3	2	3	1			1	28
03 05	35	48	27	17	14	9	6	8	2	1	5	172
06 08	16	36	19	21	5	9	1	3	1	4	2	117
09 11	3	19	9	10	4	3	1	1		1	1	52
12 14		1			1		1					3
15 16		1		1								2
N	57	122	63	58	29	26	16	13	4	9	11	408

Table 45. W_1—ΔP_b

ΔP_b (−*mb*)	*00* *19*	*20* *39*	*40* *59*	*60* *79*	*80* *99*	*100* *119*	*120* *139*	*140* *159*	*160* *179*	*180* *199*	*200* *246*	N
W_1 (mps)												
00	16	22	16	16	7	6	2		1	1	1	88
01 02	24	34	26	23	19	22	10	4	3		4	169
03 05	60	107	87	70	58	40	20	9	8	4	7	470
06 08	33	45	30	22	18	6	6	3	2			165
09 11	16	15	7	11	3	3	3	1	1			60
12 14	4		2				1					7
15 16		1	1		1							3
N	153	224	169	142	106	77	42	17	15	5	12	962

Table 46. W_1—ΔP_1

ΔP_1 ($-mb$)	*00* *29*	*30* *59*	*60* *89*	*90* *119*	*120* *149*	*150* *179*	*180* *209*	*210* *239*	*240* *269*	*270* *299*	*300* *302*	N
W_1 (mps)												
00	55	21	14	1	3		2			2		98
01 02	83	49	11	4	4							151
03 05	321	153	37	24	9	3	3	1	1	1		553
06 08	94	60	21	11	1	2	2	1			1	193
09 11	24	31	12	4			1	1				73
12 14	3	2	1		1							7
15 16	2		1									3
N	582	316	97	44	18	5	8	3	1	3	1	1078

Table 47. P_1—ΔT_a

ΔT_a (°C)	*+2·0* *+0·1*	*0·0* *−1·9*	*−2·0* *−3·9*	*−4·0* *−5·9*	*−6·0* *−7·9*	*−8·0* *−9·9*	*−10·0* *−11·9*	*−12·0* *−13·9*	*−14·0* *−15·9*	*−16·0* *−16·6*	N
P_1 (mb)											
970 979							1				1
980 989		2	3	4		1					10
990 999		16	22	10	5	2	2	1			58
1000 1009	8	58	60	21	13	4	6	1		1	172
1010 1019	3	50	33	22	6	7	2	2			125
1020 1029	2	14	9	9	5	3	1				43
1030 1039											
1040 1049											
N	13	140	127	66	29	17	12	4		1	409

Table 48. P_1—ΔT_b

ΔT_b (+°C)	*33·9* *30·0*	*29·9* *26·0*	*25·9* *22·0*	*21·9* *18·0*	*17·9* *14·0*	*13·9* *10·0*	*9·9* *6·0*	*5·9* *2·0*	*1·9* *0·0*	N
P_1 (mb)										
970 979										
980 989				1	3	4	1	5		14
990 999		1	1	2	6	8	18	29	20	85
1000 1009		1	3	13	41	56	74	92	22	302
1010 1019	1	1	4	23	65	81	86	79	24	364
1020 1029			3	22	24	46	29	42	3	169
1030 1039			1	13	10	5	1	1		31
1040 1049				1						1
N	1	3	12	75	149	200	209	248	69	966

Suppl. Part. I 1956

Table 49. P_1—ΔT_c

ΔT_c (°C)	*+5·9 +2·0*	*+1·9 −1·9*	*−2·0 −5·9*	*−6·0 −9·9*	*−10·0 −13·9*	N
P_1 (−mb)						
970 979			1			1
980 989		5	7			12
990 999	7	43	27	6	2	85
1000 1009	19	157	79	18	4	277
1010 1019	17	207	111	21	1	357
1020 1029	7	78	57	18		160
1030 1039	2	11	14	3		30
1040 1049						
N	52	501	296	66	7	922

Table 50. P_1—ΔP_a

ΔP_a (−*mb*)	*00 19*	*20 39*	*40 59*	*60 79*	*80 99*	*100 119*	*120 139*	*140 159*	*160 179*	*180 199*	*200 292*	N
P_1 (mb)												
970 979										1		1
980 989		2	3	3	1	1						10
990 999	6	15	10	11	4	6		2	1	1	2	58
1000 1009	26	60	21	20	14	11	5	7	1	2	5	172
1010 1019	21	35	25	15	9	5	4	4	1	3	3	125
1020 1029	4	10	5	9	1	3	7		1	2	1	43
N	57	122	64	58	29	26	16	13	4	9	11	409

Table 51. P_1—ΔP_b

ΔP_b (−*mb*)	*00 19*	*20 39*	*40 59*	*60 79*	*80 99*	*100 119*	*120 139*	*140 159*	*160 179*	*180 199*	*200 246*	N
P_1 (mb)												
970 979												
980 989	3	2	1	2	1	1		1	2		1	14
990 999	20	21	14	10	4	8	6				2	85
1000 1009	45	86	57	36	27	20	14	3	6	3	5	302
1010 1019	61	82	63	59	37	31	13	11	2	2	3	364
1020 1029	24	26	31	29	28	17	9	2	2		1	169
1030 1039	2	7	4	6	9	1			2			31
1040 1041									1			1
N	155	224	170	142	106	78	42	17	15	5	12	966

Table 52. P_1—ΔP_c

ΔP_c (−*mb*)	*00* *99*	*100* *199*	*200* *299*	*300* *437*	N
P_1 (mb)					
970 979		1			1
980 989	6	4	2		12
990 999	29	34	21	1	85
1000 1009	97	110	57	13	277
1010 1019	112	158	74	13	357
1020 1029	42	74	42	2	160
1030 1039	2	18	9	1	30
1040 1041					
N	288	399	205	30	922

Table 53. RH_1—ΔT_a

ΔT_a (°C)	*+2·0* *+0·1*	*0·0* *−1·9*	*−2·0* *−3·9*	*−4·0* *−5·9*	*−6·0* *−7·9*	*−8·0* *−9·9*	*−10·0* *−11·9*	*−12·0* *−13·9*	*−14·0* *−15·9*	*−16·0* *−17·9*	N
RH_1 (%)											
20 29											
30 39											
40 49	1	3	2								6
50 59	1	6	2	1		1	1				12
60 69	5	23	6	4	3			1		1	43
70 79	3	36	30	15	9	4	1	1			99
80 89	3	47	51	25	10	7	6	1			150
90 99		22	36	21	7	5	3	1			95
N	13	137	127	66	29	17	11	4		1	405

Table 54. RH_1—ΔT_b

ΔT_b (+°C)	*33·9* *30·0*	*29·9* *26·0*	*25·9* *22·0*	*21·9* *18·0*	*17·9* *14·0*	*13·9* *10·0*	*9·9* *6·0*	*5·9* *2·0*	*1·9* *0·0*	N
RH_1 (%)										
20 29						1				1
30 39				1	2	7	4	1		15
40 49				1	6	16	11	2		36
50 59				5	22	24	24	18	1	94
60 69			1	8	20	49	52	28	3	161
70 79			2	13	31	51	60	64	17	238
80 89					2	10	38	81	29	160
90 99						1	8	51	19	79
N			3	28	83	159	197	245	69	784

Table 55. RH_1—ΔP_a

ΔP_a (−*mb*)	*00* *19*	*20* *39*	*40* *59*	*60* *79*	*80* *99*	*100* *119*	*120* *139*	*140* *159*	*160* *179*	*180* *199*	*200* *292*	N
RH_1 (%)												
20 29												
30 39												
40 49	3	1		1							1	6
50 59	1	4	2		2		1	1		1		12
60 69	6	18	8	3	2	2	2				2	43
70 79	11	35	15	14	6	9	3	2	1	2	1	99
80 89	22	39	21	25	10	10	7	8	1	2	5	150
90 99	14	23	17	15	9	5	3	2	2	3	2	95
N	57	120	63	58	29	26	16	13	4	8	11	405

Table 56. RH_1—ΔP_b

ΔP_b (−*mb*)	*00* *19*	*20* *39*	*40* *59*	*60* *79*	*80* *99*	*100* *119*	*120* *139*	*140* *159*	*160* *179*	*180* *199*	*200* *246*	N
RH_1 (%)												
20 29	1											1
30 39	3	2	5	2	1	1				1		15
40 49	3	7	5	7	6	3	3		1	1		36
50 59	14	17	16	17	7	13	8	1			1	94
60 69	19	39	30	21	24	12	9	3	1	1	2	161
70 79	37	58	41	34	26	21	6	4	7	1	3	238
80 89	38	51	33	22	11	3	1	1				160
90 99	29	30	9	7	3			1				79
N	144	204	139	110	78	53	27	10	9	4	6	784

Table 57. ΔT_a—ΔT_b

ΔT_b (+°C)	*33·9* *30·0*	*29·9* *26·0*	*25·9* *22·0*	*21·9* *18·0*	*17·9* *14·0*	*13·9* *10·0*	*9·9* *6·0*	*5·9* *2·0*	*1·9* *0·0*	N
ΔT_a (°C)										
+2·0 +0·1				1		2	8	2		13
0·0 −1·9					6	15	35	60	11	127
−2·0 −3·9					1	9	25	59	17	111
−4·0 −5·9					2	1	9	31	11	54
−6·0 −7·9						1	5	12	5	23
−8·0 −9·9							2	7	2	11
−10·0 −11·9							1	5	2	8
−12·0 −13·9								1	2	3
−14·0 −15·9										
−16·0 −17·9									1	1
N				1	9	28	85	177	51	351

Table 58. ΔT_a—ΔP_a

ΔP_a (−mb)	*00 19*	*20 39*	*40 59*	*60 79*	*80 99*	*100 119*	*120 139*	*140 159*	*160 179*	*180 199*	*200 292*	N
ΔT_a (°C)												
+2·0 +0·1	4	5	1	1	1						1	13
0·0 −1·9	45	68	17	8		1		1				140
−2·0 −3·9	8	45	37	21	9	4		2			1	127
−4·0 −5·9		3	8	23	13	15	3	1				66
−6·0 −7·9		1		4	4	4	10	3	2	1		29
−8·0 −9·9			1	1	1	2	2	4	1	3	2	17
−10·0 −11·9					1		1	2	1	4	3	12
−12·0 −13·9										1	3	4
−14·0 −15·9												
−16·0 −17·9											1	1
N	57	122	64	58	29	26	16	13	4	9	11	409

Table 59. ΔT_a—ΔP_b

ΔP_b (−mb)	*00 19*	*20 39*	*40 59*	*60 79*	*80 99*	*100 119*	*120 139*	*140 159*	*160 179*	*180 199*	*200 246*	N
ΔT_a (°C)												
+2·0 +0·1	2	3	4	3	1							13
0·0 −1·9	25	38	26	18	7	5	1	4	2		1	127
−2·0 −3·9	28	38	20	13	7		3			1	1	111
−4·0 −5·9	19	18	6	6	2	2	1					54
−6·0 −7·9	7	11	3	2								23
−8·0 −9·9	5	4	1	1								11
−10·0 −11·9	6	2										8
−12·0 −13·9	1		1	1								3
−14·0 −15·9												
−16·0 −17·9		1										1
N	93	115	61	44	17	7	5	4	2	1	2	351

Table 60. ΔT_b—ΔT_c

ΔT_c (°C)	*+5·9 +2·0*	*+1·9 −1·9*	*−2·0 −5·9*	*−6·0 −9·9*	*10·0 13·9*	N
ΔT_b (+°C)						
33·9 30·0						
29·9 26·0		1	1		1	3
25·9 22·0		5	4	1		10
21·9 18·0	3	36	26	2	1	68
17·9 14·0	7	78	39	3	1	128
13·9 10·0	7	113	53	12	1	186
9·9 6·0	18	101	51	11	1	182
5·9 2·0	13	94	77	19		203
1·9 0·0	1	17	14	6	1	39
N	49	445	265	54	6	819

Suppl. Part. I 1956

Table 61. ΔT_b—ΔP_a

ΔP_a (−mb)	*00* *19*	*20* *39*	*40* *59*	*60* *79*	*80* *99*	*100* *119*	*120* *139*	*140* *159*	*160* *179*	*180* *199*	*200* *292*	N
ΔT_b (+°C)												
33·9 30·0												
29·9 26·0												
25·9 22·0												
21·9 18·0			1									1
17·9 14·0		4	2	1		2						9
13·9 10·0	4	11	9	3				1				28
9·9 6·0	11	33	17	8	4	2	5	4			1	85
5·9 2·0	30	49	26	28	14	13	4	3	4	4	2	177
1·9 0·0	7	12	3	11	5	3	3			3	4	51
N	52	109	58	51	33	20	12	8	4	7	7	351

Table 62. ΔT_b—ΔP_b

ΔP_b (−mb)	*00* *19*	*20* *39*	*40* *59*	*60* *79*	*80* *99*	*100* *119*	*120* *139*	*140* *159*	*160* *179*	*180* *199*	*200* *246*	N
ΔT_b (+°C)												
33·9 30·0										1		1
29·9 26·0							1		1		1	3
25·9 22·0						2	3	1	2		4	12
21·9 18·0	1	6	5	10	24	15	6	2	4	2		75
17·9 14·0	4	18	30	33	24	14	14	4	4		4	149
13·9 10·0	16	32	36	45	29	25	8	4	2		3	200
9·9 6·0	24	58	47	28	19	19	6	5	2	1		209
5·9 2·0	80	85	42	22	10	3	4	1		1		248
1·9 0·0	30	25	10	4								69
N	155	224	170	142	106	78	42	17	15	5	12	966

Table 63. Mean Annual Lapse Rates by Type

Units		*A layer*			*B layer*			*C layer*		
		$\overline{\Delta T}$	$\overline{\Delta H}$	$\frac{\overline{\Delta T}}{\overline{\Delta H}}$	$\overline{\Delta T}$	$\overline{\Delta H}$	$\frac{\overline{\Delta T}}{\overline{\Delta H}}$	$\overline{\Delta T}$	$\overline{\Delta H}$	$\frac{\overline{\Delta T}}{\overline{\Delta H}}$
ABC	N	°C	m	$\frac{°C}{100\ m}$	°C	m	$\frac{°C}{100\ m}$	°C	m	$\frac{°C}{100\ m}$
031	310				+12	564	+2·1	−3	1610	−·2
032	164				+13	494	+2·6	0	1060	0
131	149	−3	409	−·7	+ 5	335	+1·5	−3	1295	−·2
033	80				+12	359	+3·3	+2	1120	+·2
130	81	−4	599	−·7	+ 5	445	+1·1			
132	76	−4	479	−·8	+ 6	295	+2·0	0	699	0
030	61				+12	709	+1·7			
001	36							−4	1884	−·2
101	30	−4	604	−·7				−3	1340	−·2
133	18	−2	394	−·5	+ 6	227	+2·6	+2	1025	+·2
102	15	−5	744	−·7				0	730	0
002	11							0	1164	0

Table 64. T_1—$\overline{R_1}$ *and* $\overline{R_2}$

T_1	N	$\overline{\Delta T_b}$	$\overline{\Delta P_b}$	$\overline{R_1} = \frac{\overline{\Delta T_b}}{\overline{\Delta P_b}}$	$\overline{R_2} = \overline{R_1}\,\overline{(\Delta T)_b}$
°C		+°C	mb	$\frac{°C}{mb}$	$\frac{(°C)^2}{mb}$
+ 2	10	2·9	33	·09	·26
+ 1	28	3·2	30	·11	·35
0	49	3·2	40	·08	·26
− 1	28	3·3	38	·09	·30
− 2	16	2·7	25	·11	·30
− 3	17	3·7	32	·12	·44
− 4	17	3·0	27	·11	·33
− 5	10	3·6	34	·11	·40
− 6	13	4·3	42	·10	·43
− 7	16	4·4	40	·11	·48
− 8	16	3·5	27	·13	·46
− 9	10	4·5	31	·15	·68
−10	7	5·4	33	·16	·86
−11	14	4·3	49	·09	·39
−12	20	4·4	28	·16	·70
−13	13	7·1	46	·15	1·07
−14	17	6·6	41	·16	1·06
−15	10	4·8	36	·13	·62
−16	5	6·3	35	·18	1·13
−17	13	5·6	26	·22	1·23
−18	10	6·6	49	·13	·86
−19	6	8·5	45	·19	1·62
−20	7	7·3	54	·14	1·02
−21	13	8·4	57	·15	1·26
−22	10	6·7	58	·12	·80
−23	19	9·7	71	·14	1·36
−24	23	9·0	67	·13	1·17
−25	16	9·8	83	·12	1·18
−26	28	9·2	63	·15	1·38
−27	24	9·0	66	·14	1·26
−28	23	10·6	72	·15	1·59
−29	26	10·7	73	·15	1·61
−30	24	12·1	77	·16	1·94
−31	23	11·0	71	·15	1·65
−32	31	12·0	61	·20	2·40
−33	28	11·7	71	·16	1·87
−34	34	11·4	75	·15	1·71
−35	34	12·2	61	·20	2·44
−36	33	13·8	74	·19	2·62
−37	31	15·8	87	·18	2·84
−38	31	13·5	73	·18	2·43
−39	27	16·0	86	·19	3·04
−40	34	14·3	77	·19	2·72
−41	18	16·7	83	·20	3·34
−42	20	14·9	66	·23	3·43
−43	13	15·5	84	·18	2·79
−44	17	17·0	77	·22	3·74
−45	7	17·8	91	·20	3·56
−46	15	19·3	115	·17	3·28
−47	7	20·7	139	·15	3·11
−48	2	16·4	74	·22	3·61
−49	2	16·4	55	·30	4·92
−50	1	18·1	113	·16	2·90
−51	1	17·3	77	·22	3·81
	1097				

Table 65. Extreme and Usual Values of Variables for Entire Period

Field	*Unit*	*N*	*Abs. Max.*	*Abs. Min.*	*Most frequent values*	
					Primary	*Secondary*
T_1	°C	1097	+2	−51	+2 to −2	−35 to −39
W_1	mps	1092	16	0	03 to 05	06 to 08
P_1	mb	1097	1041	974	1000 to 1019	
RH_1	%	915	100	29	70 to 79	80 to 89
ΔT_a	°C	409	−16·6	+0·8	0·0 to −3·9	
ΔT_b		966	+30·5	+0·5	0·0 to 5·9	6·0 to 13·9
ΔT_c		922	−12·4	+5·8	+1·9 to −1·9	−2·0 to −5·9
ΔP_a	mb	409	292	7	20 to 39	40 to 59
ΔP_b		966	246	3	20 to 39	40 to 59
ΔP_c		922	437	10	100 to 199	00 to 99
R_1	°C/mb	966	2·01	·02		
R_2	$(°C)^2$/mb	966	32·9	·01	0·0 to 1·0	1·0 to 2·0

Note. Mean values by *ABC* type are given in *Tables 1* and *2*.

Table 66. Percentage Frequency of Types by Month (⩾10 per cent)

ABC	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*J*	*F*	*M*	*A*	*M*	
132	18				12							16	
131	30	32	21	21	18						12	23	
130	27	15		11									
101		11	12										
033								16	14	20			
032					12	19	16	21	27	35	20		
031				24	28	25	57	44	36	33	36	23	
030						18							
001			13	13									
000			16										
N	77	110	86	94	50	93	112	110	92	109	108	56	1097

CURRENT STATUS OF SEA ICE RECONNAISSANCE AND FORECASTING FOR THE AMERICAN ARCTIC

DR. CHARLES C. BATES*
U.S. Navy Hydrographic Office, Washington, D.C.

In 1949 the U.S. Navy began a steadily developing program for observing and forecasting sea ice conditions throughout the American Arctic. In order to illustrate the present state of the art, a résumé is given of a typical monthly briefing of the Commander, Military Sea Transport Service, and his staff concerning the existing and probable future ice conditions to be encountered during the annual resupply of northern bases. Scattered ship or shore reports are not an adequate basis for stating whether ice conditions were "light" or "heavy" within a given basin. In order to provide the truly synoptic picture of ice conditions existing in the Arctic between the Bering Sea and Greenland and Newfoundland, a reporting system based on shore and ship observations has been combined with an extensive aerial reconnaissance program which now involves flying over 500,000 nautical miles annually. This program utilizes the services of many polar activities, including the Air Weather Service of the U.S. Air Force, the Royal Canadian Navy and Air Force, the Meteorological Service of the Canadian Department of Transport, the Defence Research Board of Canada, the U.S. Weather Bureau, and the U.S. Coast Guard. Most ice observations are primarily visual and thus semi-quantitative; accordingly, development work on ice physics and time-lapse photography is also conducted. Freeze-up forecasting, being based on observed meteorological and oceanographic conditions, is quantitative. However, metamorphic and advective processes operating in the ice are less well understood, as are the controls operating in the Western Hemisphere's six major centers of ice disintegration. Verification values for the 48-hr, 5-day, 30-day ice forecasts and the seasonal outlooks (120–180 days) are given, and a comparison is made between the accuracy of seasonal outlooks and ice atlases. Research projects now under way to provide more quantitative methods include radio-monitored weather buoys placed on the polar pack north of Alaska and the potential programming of high-speed computers to determine movement of icefields under varying fields of wind and current.

uppl.
art.I
956

PROBABLY the major influence of the polar atmosphere upon naval operations is that the temperature régime causes much of the Arctic to be covered with water in the solid, rather than the liquid, state. This "pack ice" is made up of two major types, one of which, being quasi-stationary, forms and disintegrates largely *in situ*, while the other, after formation, drifts southward as a steady stream to disintegrate in lower latitudes. The area covered by ice of both types in the American Arctic is shown in *Fig. 1.* Because of this ice cover, the supply of the numerous northern bases by ships is crowded into a few months each year, thereby making scheduling and implementation of this supply operation a major problem to the United States Navy's

* The opinions expressed in this paper are those of the author as an individual and are not to be construed as necessarily reflecting the official views of the Navy Department.

THE ARCTIC REGIONS

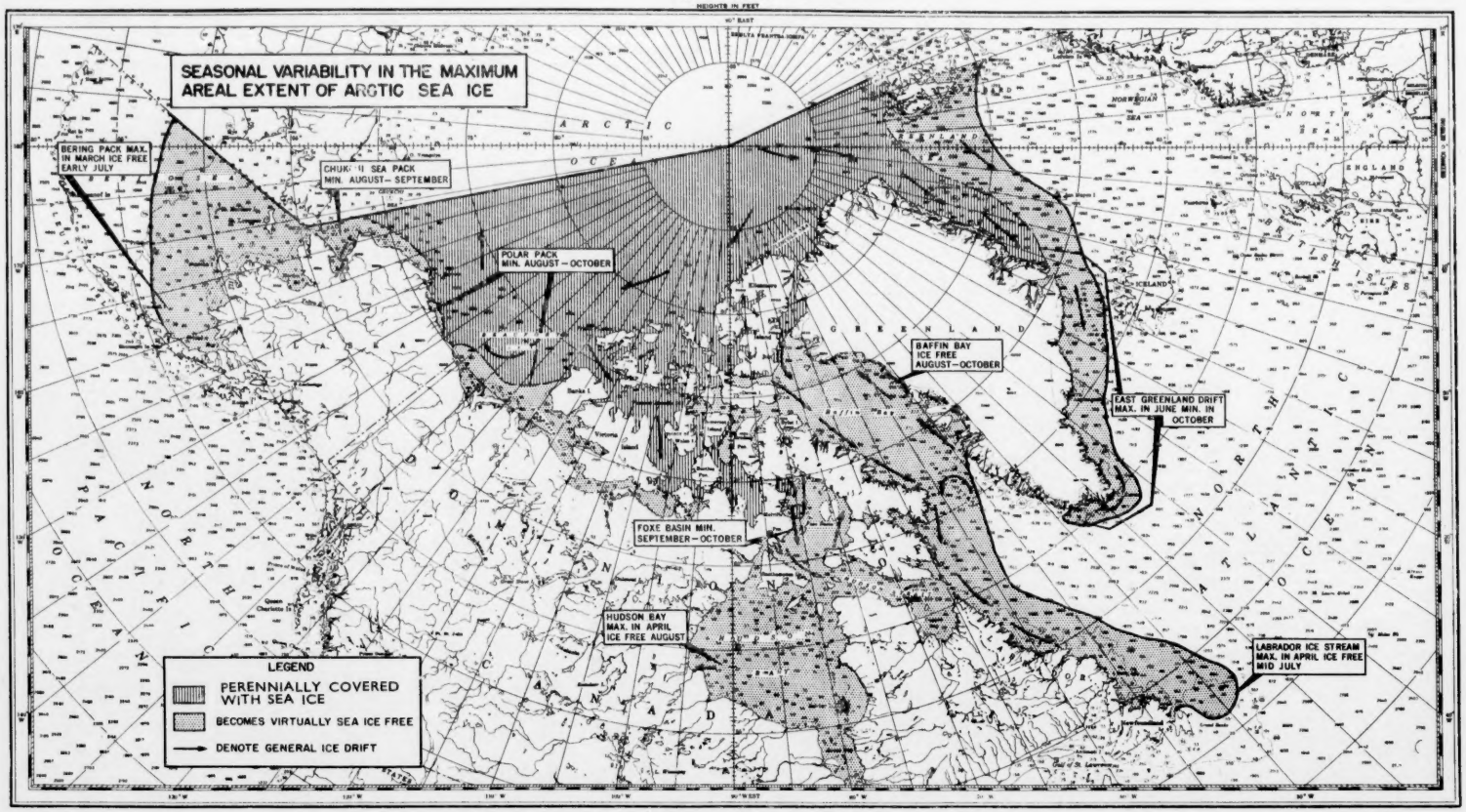

SOUNDINGS IN FATHOMS
HEIGHTS IN FEET
SEASONAL VARIABILITY IN THE MAXIMUM AREAL EXTENT OF ARCTIC SEA ICE
BERING PACK MAX. IN MARCH ICE FREE EARLY JULY
CHUKCHI SEA PACK MIN. AUGUST–SEPTEMBER
POLAR PACK MIN. AUGUST–OCTOBER
BAFFIN BAY ICE FREE AUGUST–OCTOBER
EAST GREENLAND DRIFT MAX. IN JUNE MIN. IN OCTOBER
FOXE BASIN MIN. SEPTEMBER–OCTOBER
HUDSON BAY MAX. IN APRIL ICE FREE AUGUST
LABRADOR ICE STREAM MAX. IN APRIL ICE FREE MID JULY
LEGEND
PERENNIALLY COVERED WITH SEA ICE
BECOMES VIRTUALLY SEA ICE FREE
DENOTE GENERAL ICE DRIFT
ARCTIC OCEAN
NORTH ATLANTIC OCEAN
GREENLAND
ICELAND
NORWEGIAN SEA
BRITISH ISLES
ENGLAND
CANADA
ALASKA

Fig. 1. Maximal and minimal limits of sea-ice in the American Arctic

Supp
Par
195

Suppl.
Part. I
1956

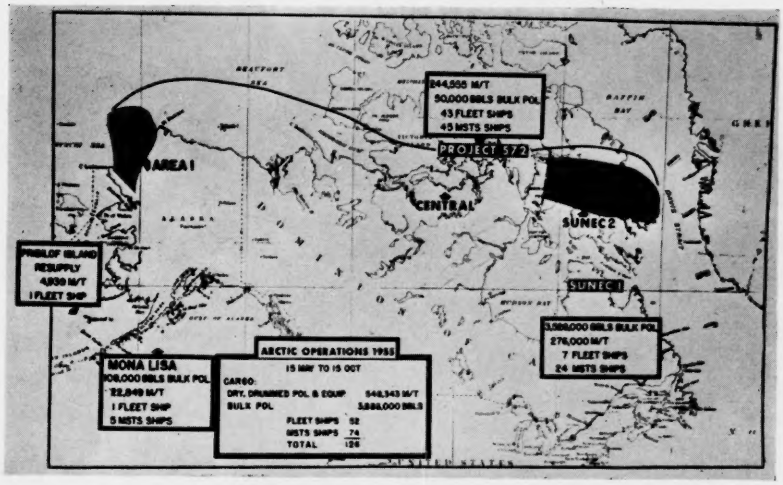


Fig. 2. Ships and tonnage involved in the 1955 Arctic operations of the Military Sea Transport Service

(*Facing page* 286)

Sup
Par
195

Military Sea Transport Services (MSTS), a post-World War II agency. MSTS's first major polar operation was the transfer of men and supplies to Thule, Greenland, in 1951 for the purpose of establishing that large air-base 750 nautical miles from the North Pole. Subsequent years brought about the establishment of smaller bases along the coast of Labrador and northern Alaska. In 1955 came the establishment of the vast net of radar stations termed the Distant Early Warning (DEW) Line along latitude 72°N

Suppl.
Part.I
1956

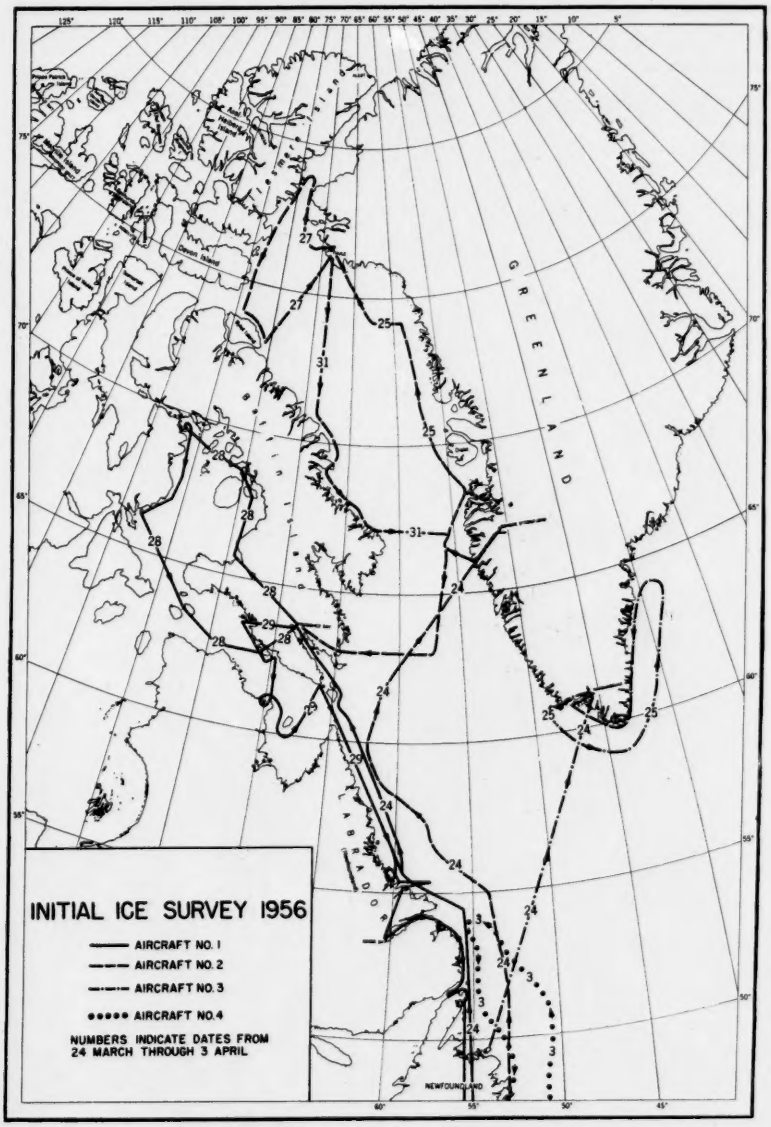


Fig. 3. Tracks of ice reconnaissance aircraft in late March and early April 1956

from Alaska to Baffin Bay, a distance of roughly 3600 miles. Because of this steady growth, the 1955 Arctic supply mission required 126 Canadian Navy, United States Navy and Coast Guard and merchant ships to move 500,000 tons of measured cargo and 3,500,000 barrels of petroleum products (see *Fig. 2*). Such vast operation requires prompt and accurate ice observations and forecasts, and it is the purpose of this paper to relate how this requirement is currently being met.

In order to obtain a proper understanding of the steps taken to provide such an ice service, imagine yourself in MSTS Headquarters, Washington,

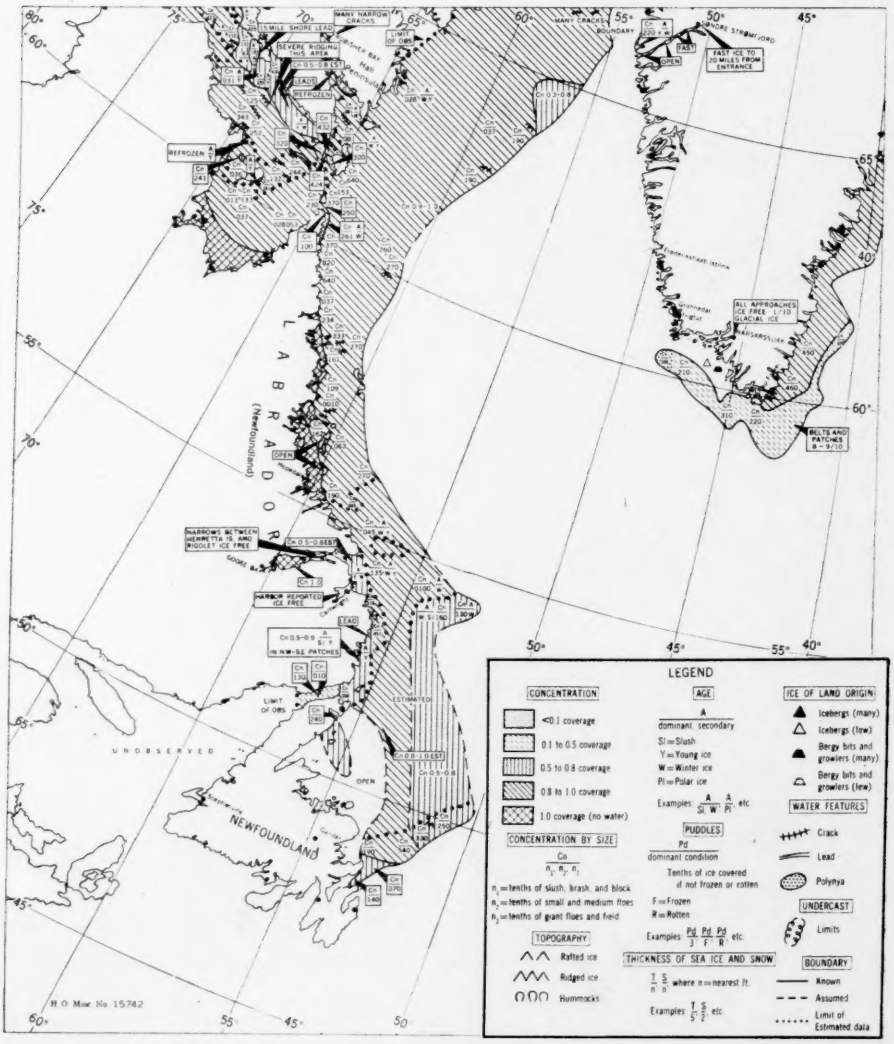


Fig. 4. Ice conditions observed between Newfoundland and Davis Strait 24 March–4 April 1956

D.C., on 17 April 1956. On this date the comprehensive monthly ice briefing is being given to Vice-Admiral Francis D. Denebrink, U.S.N., the commander, and his staff. Because eastern operations start at an earlier date, the area covered is the region between Newfoundland and north-west Greenland. Outlines of expected conditions through mid-August 1956 are presented. The briefer is W. Wittman, a senior ice forecaster who spent the

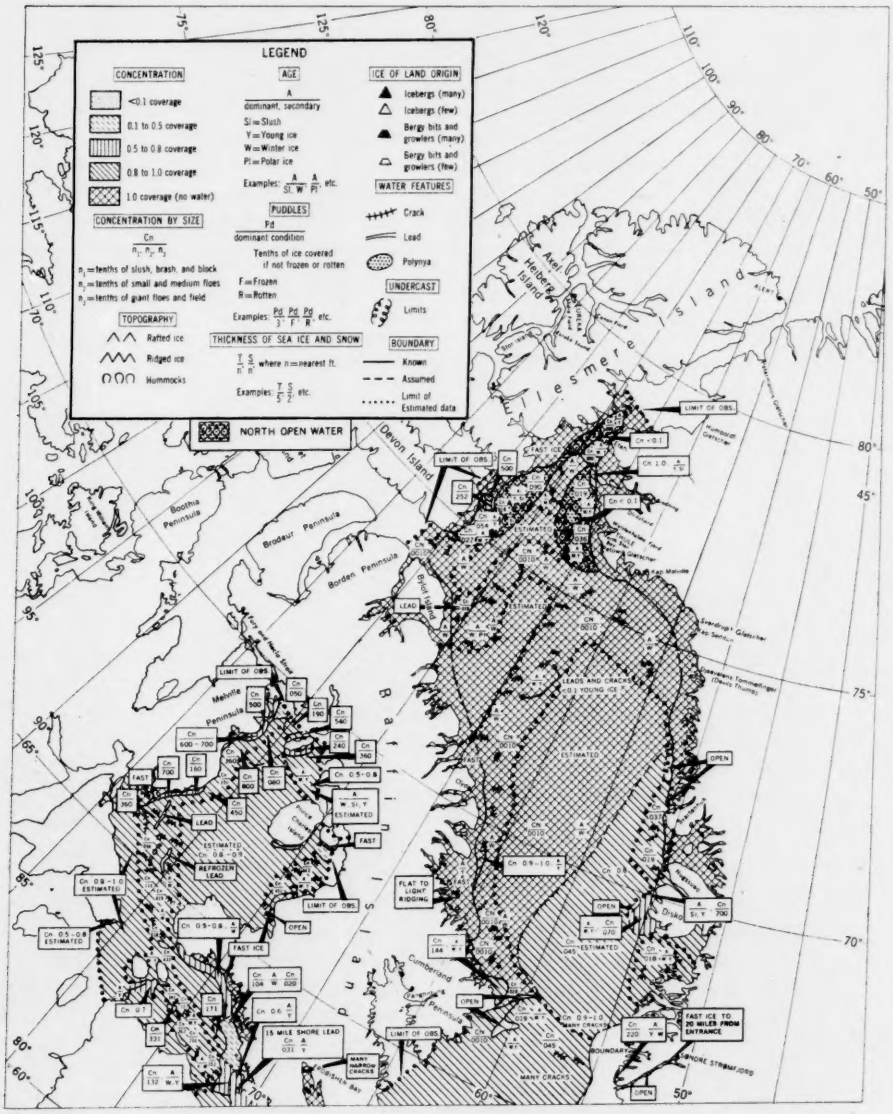


Fig. 5. Ice conditions observed north of Hudson Strait, 24 March–4 April 1956

Suppl. Part. I 1956

period of 24 March–3 April 1956 with G. MacDowell, trainee ice forecaster, and four Navy enlisted ice observers flying over the area in P2V reconnaissance aircraft (see *Fig. 3*). Upon completion of this early reconnaissance with the results shown in *Figs. 4* and *5*, the staff of the Experimental Oceanographic Forecasting Central of the U.S. Navy Hydrographic Office proceeded through all the varied steps required in making a long-range ice outlook (see *Table 1*) and came to the following conclusions:

(*a*) Comparison of ice boundaries existing in late March of 1954, 1955 and 1956 (see *Fig. 6*) indicated that less ice was present in the lower portion of the Labrador Drift Stream than in either 1954 or 1955. Furthermore, off the coast of Newfoundland, an almost ice-free shore lead stretched southward from Belle Isle Strait to the proximity of Trinity Bay. This was not evident either in 1955 or 1956. Further northward, from the latitude of Belle Isle to the Greenland coast just south of Egedesminde, the ice boundary occupied a very similar position to that held in 1954 and 1955. The storis boundary off southern Greenland was very similar in configuration to that observed

Table I. Steps in Preparation of Long-Range Ice Outlook

(*a*) Ice Potential Computations
  1. Theoretical Initial Ice Formation
  2. Theoretical Ice Growth

(*b*) Analysis of Recent Climatology
  1. Air Temperature Anomalies
  2. Degree-Day Accumulations
  3. Snow Cover
  4. Storm Frequency and Tracks

(*c*) Analysis of Historical Ice Data

(*d*) Analysis of Recent Ice Formation
  1. Late Fall Ice Reconnaissance
  2. Ice Thickness Data
  3. Theoretical Ice Drift
  4. Early Ice Reconnaissance

in 1955, although the quantity of ice was somewhat less. Between Kaps Thorvaldson and Farvel, a shore lead was in evidence unlike 1955. In Hudson Strait, the Big Island lead was wider than in 1955; however, no narrow lead existed between Cape Dorset and Markham Bay as it did during 1955. A stronger north-east to south-west component of ice drift in Hudson Strait than in 1955 was suggested by a comparatively greater quantity of pressure-ice along the south-western shore. In the Melville Bugt approaches to Thule, no extensive fast-ice existed as was the case in 1954. The 1956 fast-ice boundary closely approximated that of 1955, but few leads comparable to those witnessed the previous year were to be found. Although the outer ice limits of the North Open Water were very similar to those observed in 1954 and 1955, a wide tongue of winter ice stretched northward from the line Thule-Philpots Island, dividing the open water and the young ice usually characteristic of this remarkable area. This condition indicates colder water temperatures than usual and delayed ice disintegration for the Melville Bugt area.

(*b*) Snow cover (see *Fig. 7*) was very heavy during the winter of 1955–56 along the entire Labrador coast and in lower Ungava Bay, 25 in.

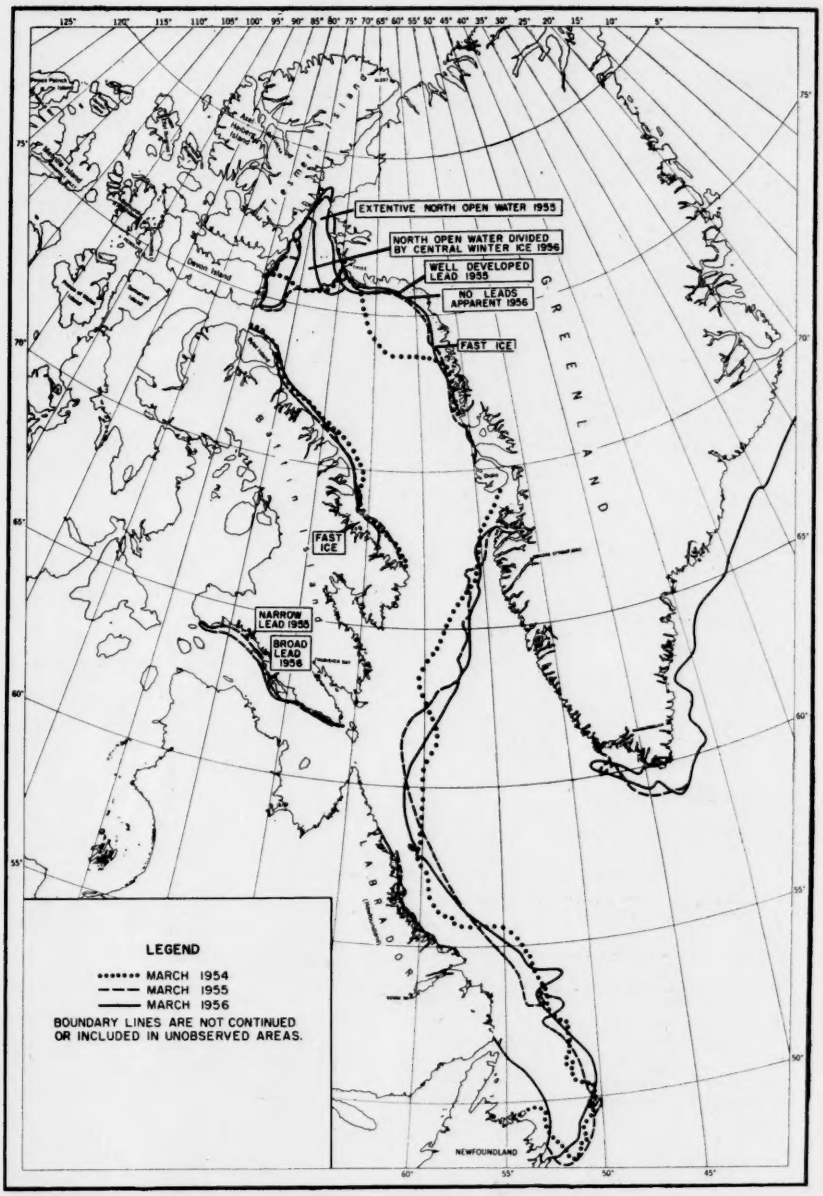


Fig. 6. Ice boundaries between Newfoundland and Smith Sound during March of 1954, 1955 and 1956

more snow falling in the Hamilton Inlet and Fort Chimo regions than the previous season. As a result, winter ice in the offshore portions of the pack is expected to be thinner than normal.

(*c*) Air temperatures in Foxe Basin, Hudson Strait and along Labrador

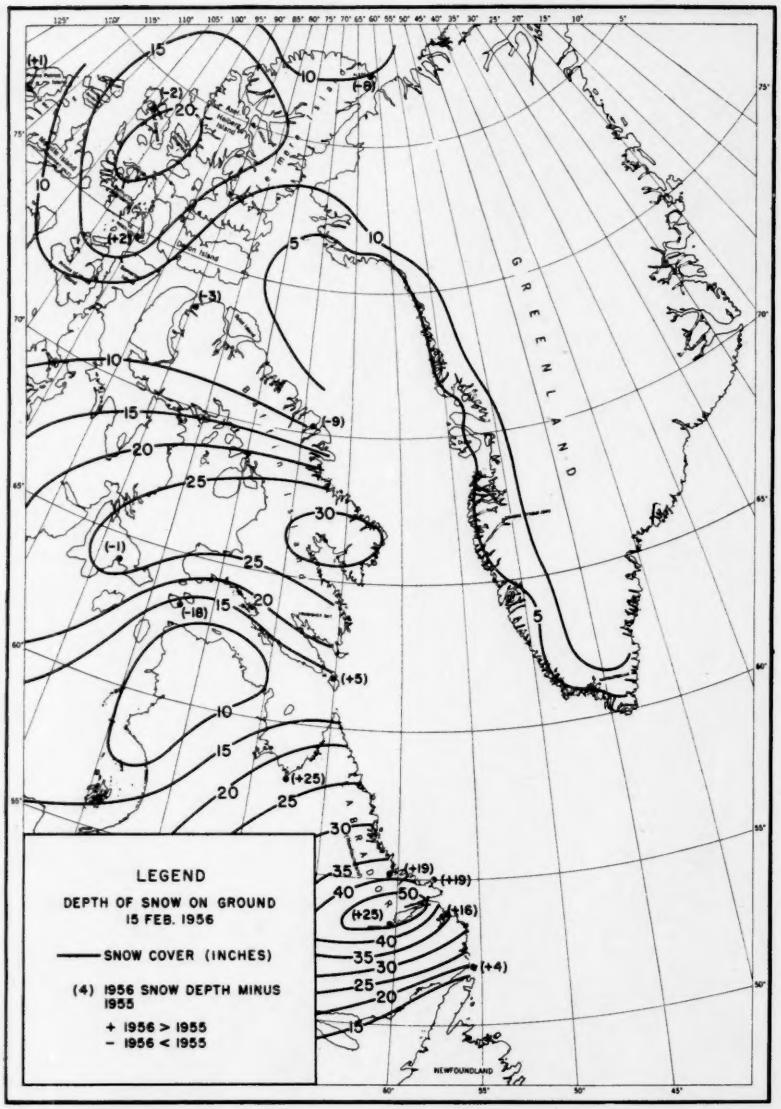


Fig. 7. Depth of ground snow cover in eastern American Arctic on 15 February 1956

have been considerably warmer during the winter of 1955–56 than normally (see *Figs. 8* and *9*). Degree-days in the latter illustration are computed by using 32°F as the base. The deficiency of 1000 degree-days from November 1955 through February 1956 over the previous season's degree-day count (see *Fig. 10*) is roughly equivalent to an ice growth of 12 in. less. Elsewhere in the area, near normal

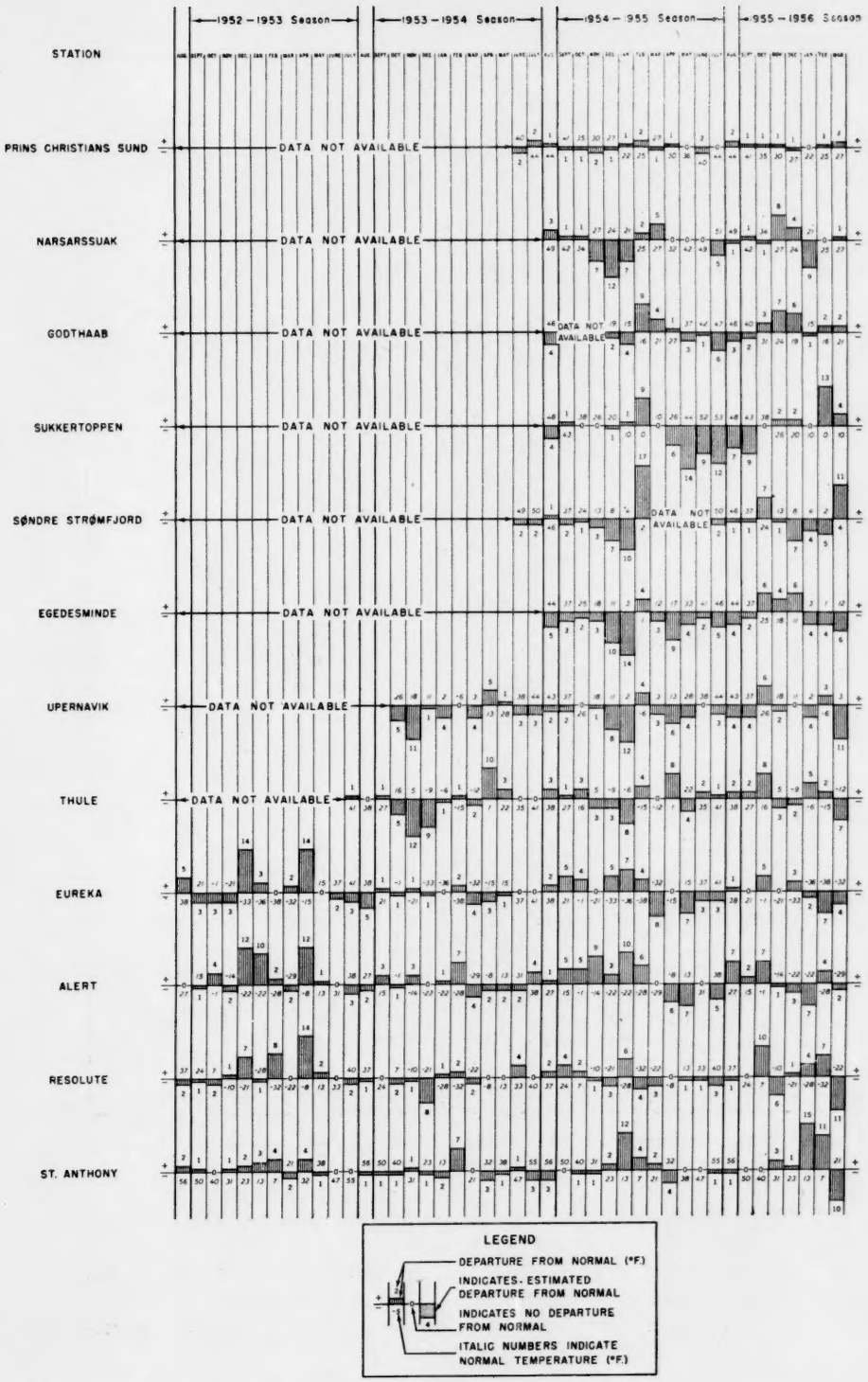


Fig. 8. Mean monthly surface air temperatures and deviation from normal for eastern American Arctic land stations, 1952–56

Suppl.
Part. I
1956

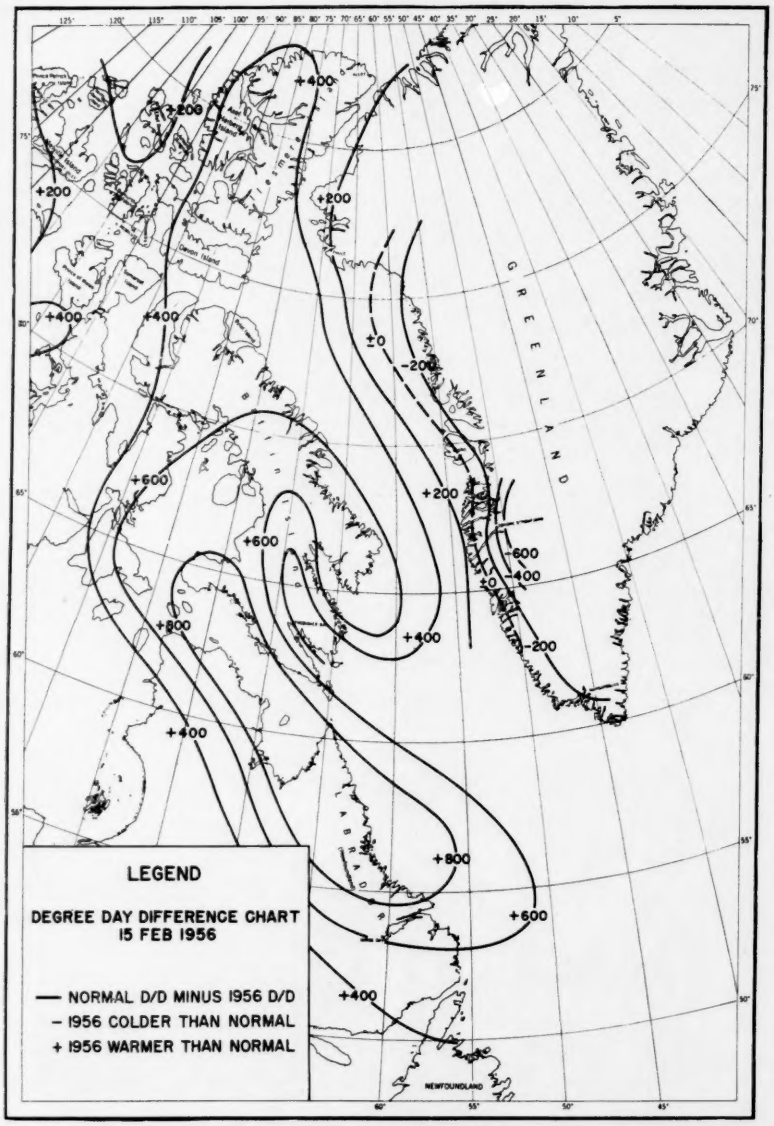


Fig. 9. Degree-day chart (base of 32°F) showing winter anomalies from normal as of 15 February 1956 for the eastern American Arctic

temperatures prevailed except for Sondre Stromfjord where considerably colder temperatures prevailed.

(*d*) Theoretical ice drift vectors indicate that the ice of Melville Bugt was less subject to divergent drift in 1956 than during the previous year, implying a slower opening to navigation. Along the coast of Baffin

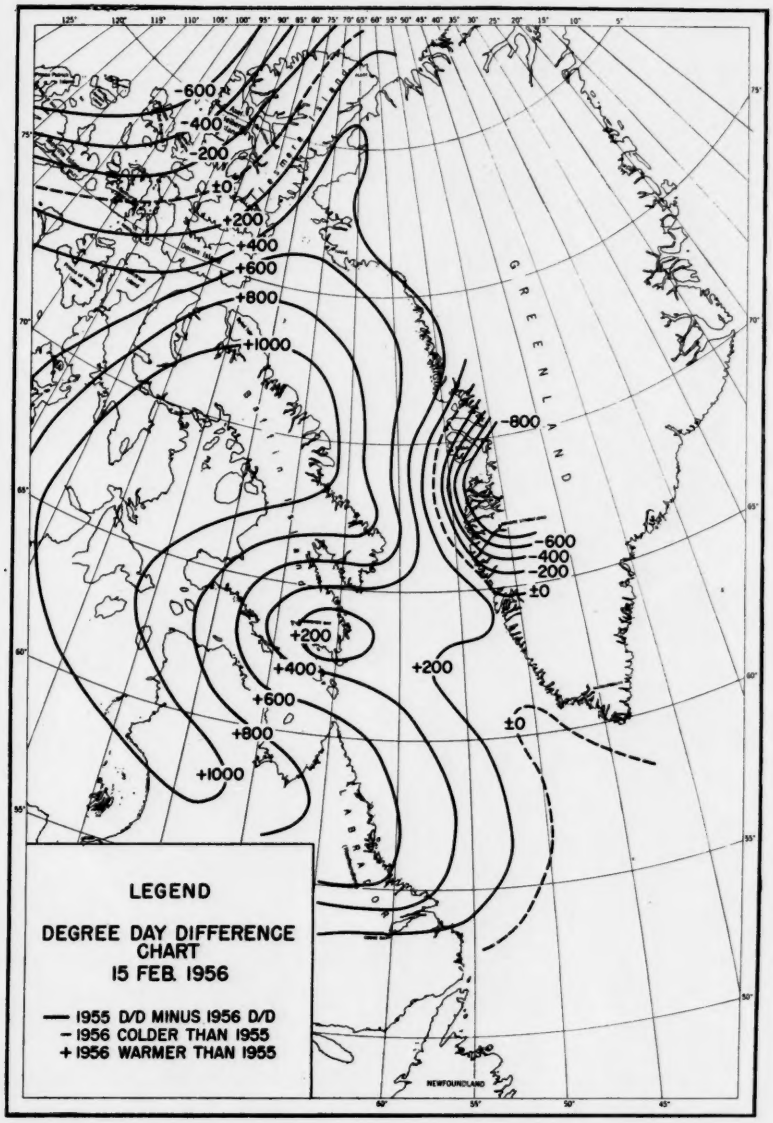
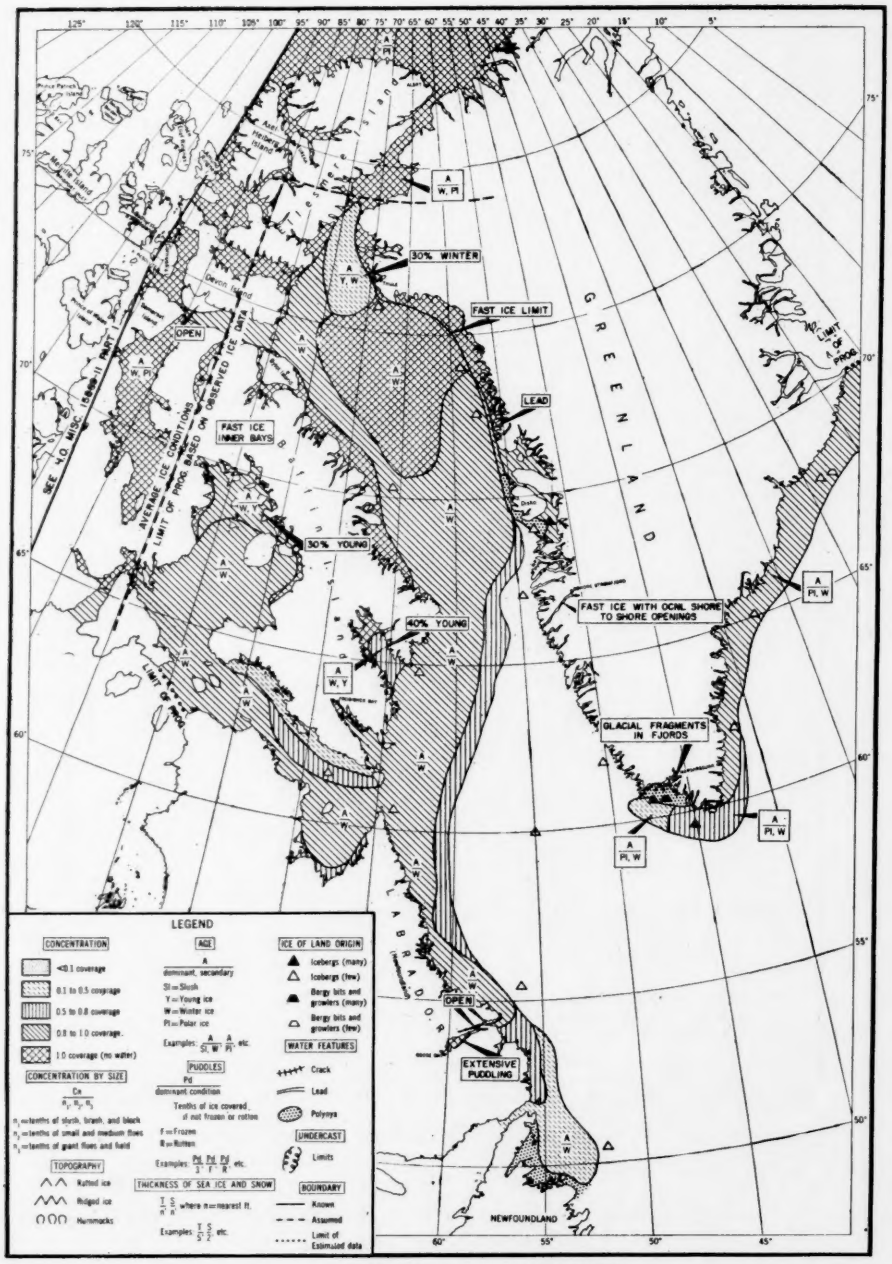


Fig. 10. Degree-day chart (base of 32°F) showing difference in 1955 and 1956 winter totals up to 15 February

Island and within the Labrador Drift Stream the north-to-south component of theoretical ice movement from December through February was almost a nautical mile per day less in 1956 than in 1955. This should make for somewhat improved ice conditions south of Davis Strait, there being lesser amounts of heavy winter ice feeding out of

Suppl.
Part.I
1956



PROGNOSTIC ICE CHART, MID-MAY 1956

Fig. 11. Forty-day ice forecast for mid-May 1956 in the eastern American Arctic

Baffin Bay. In the southern Labrador area, drift-ice will also be exposed per unit of travel distance to disintegration processes for a longer period of time, resulting in an accelerated northward recession of the southern ice limit during June and July. While Hudson Strait had comparable theoretical ice-drift characteristics in 1955 and 1956, Foxe Basin reflected an earlier opening in the northern extremities as a result of a more persistent offshore drift. In fact, this drift had opened a 17 by 100-mile lead there during late March even though temperatures had previously averaged below −30°F for the previous 3 months.

(*e*) The 40-day long-range forecast can be portrayed graphically in *Fig. 11.* A most important aspect of this chart is that, in general, there is 5/10 to 8/10 concentration of storis in the direct approaches to Simutak and Narsarssuak, but a definite lesser concentration to the northwestward, thereby making the preferred route of approach from Kap Thorvaldsen and Kap Desolation Island southward along the coastline into Scov-Fjord or Breve Fjord. Of equal importance is the scattered ice concentration of 5/10 or less to the south of Belle Isle Strait, a very optimistic condition when compared to those existing in May 1955.

Suppl.
Part. I
1956

(*f*) The 70-day ice outlook for mid-June 1956 (see *Fig. 12*) shows the boundary of the Labrador pack-ice to have receded in a normal manner so that Hamilton Inlet and coastal points to the south of the inlet are clear. On the Greenland side, though, ice conditions are actually a little worse than in May, there being a smaller but more concentrated tongue of storis between Kap Farvel and Simutak. In Baffin Bay, the North Open Water has developed in size but more slowly than in 1955, while ice concentrations off Baffin Island proper are not as great as in 1955.

Similar 100-day and 130-day outlooks were then provided in chart form for mid-July and mid-August 1956. Finally, the briefing closed with a detailed rundown port by port of the expected opening days for icebreaker escorted and for unescorted supply ships.

In forecasting fluctuations of any geophysical phenomenon, it is obviously necessary to know the exact conditions existing at the time of forecast. Prior to 1950, ice data for the American Arctic was supplied primarily by isolated ship reports. Most of these reporting ships did not penetrate deeply into the pack-ice, and their reports when compiled at the end of the supply season could give a small uncertain sample in terms of either distance or time as to what ice conditions actually existed over vast polar areas. Now 5 years of increasingly comprehensive aerial ice reconnaissance are available; it is evident that generalizations as to ice budgets or the type of shipping season, if based on the pre-1950 data, must be accepted with caution. Spatial variations of sea-ice in time can be about as great in a latitudinal as in a longitudinal sense. For example, *Fig. 13* portrays the ice boundary during mid-June 1953, 1954 and 1955 in the Labrador Sea-Davis Strait-Baffin Bay circulatory system. Marked variations are found in the tongue of ice being advected southwards along the Labrador coast, in the width of the storis

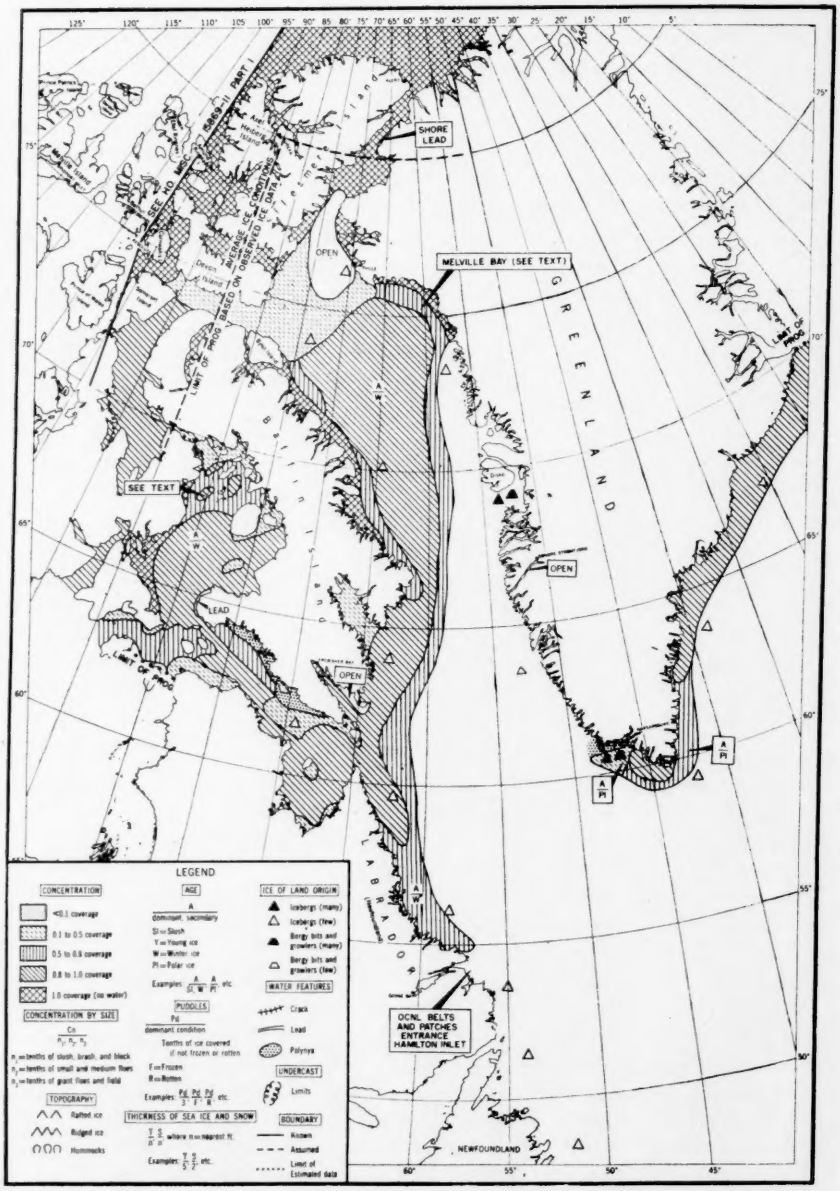


Fig. 12. Seventy-day ice forecast of ice conditions during mid-June 1956 in the eastern American Arctic

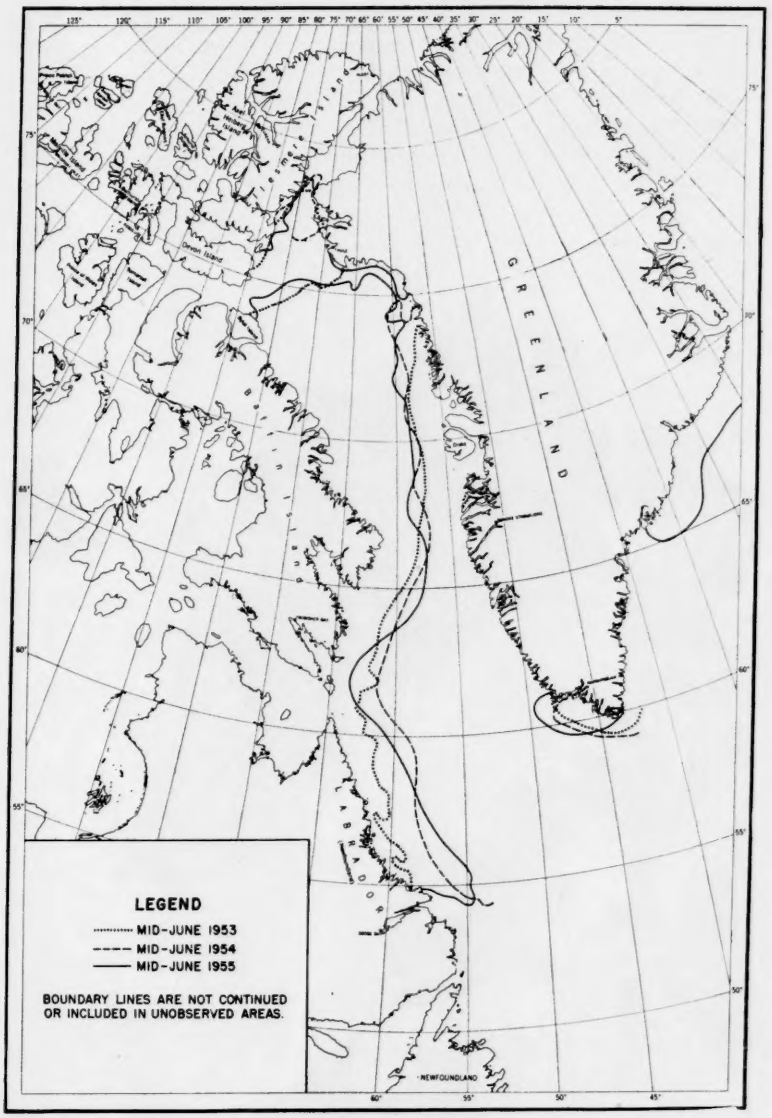


Fig. 13. Sea-ice boundaries in the eastern American Arctic during mid-June of 1953, 1954 and 1955

tongue off southern Greenland, in the south-eastern boundary of the Baffin Bay pack north-west of Disko Island, and in the areal expanse of the North Open Water off Thule, Greenland. Comparable variability existed a month later when in mid-July 1953, 1954 and 1955 (see *Fig. 14*) the tip of the Labrador pack also fluctuated in location, and there was even greater variation in the northern margin of the Baffin Bay pack. The greatest

latitudinal variations of the American Arctic are found off northern Alaska, as shown in the ice charts for early September of 1953, 1954 and 1955 (see *Fig. 15*). The year 1954 is noted as being one in which the polar pack, i.e. polar ice with a concentration of 8/10 or greater, retreated far offshore, whereas in 1955 there was no permanent retreat at all. The year 1953 can be characterized as intermediate between these extreme conditions. Fluctuation in the ice margin in the approaches to Amundsen Gulf and western

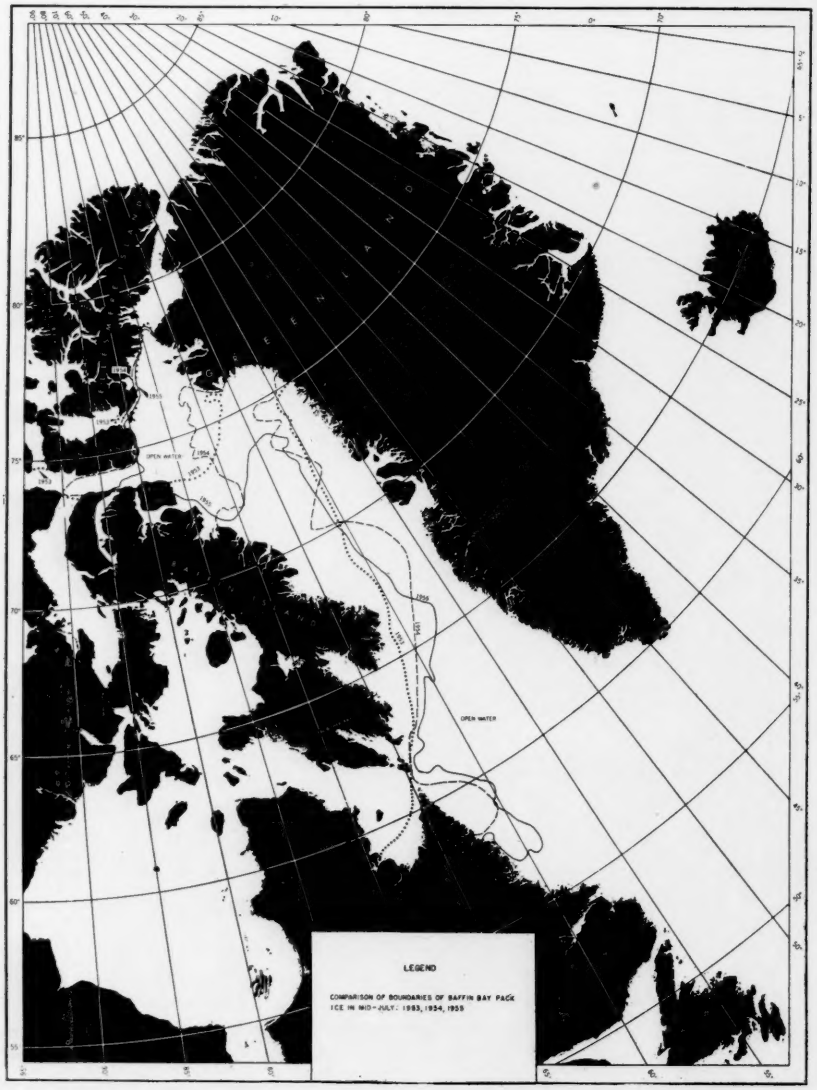


Fig. 14. Sea-ice boundaries in the eastern American Arctic during mid-July of 1953, 1954 and 1955

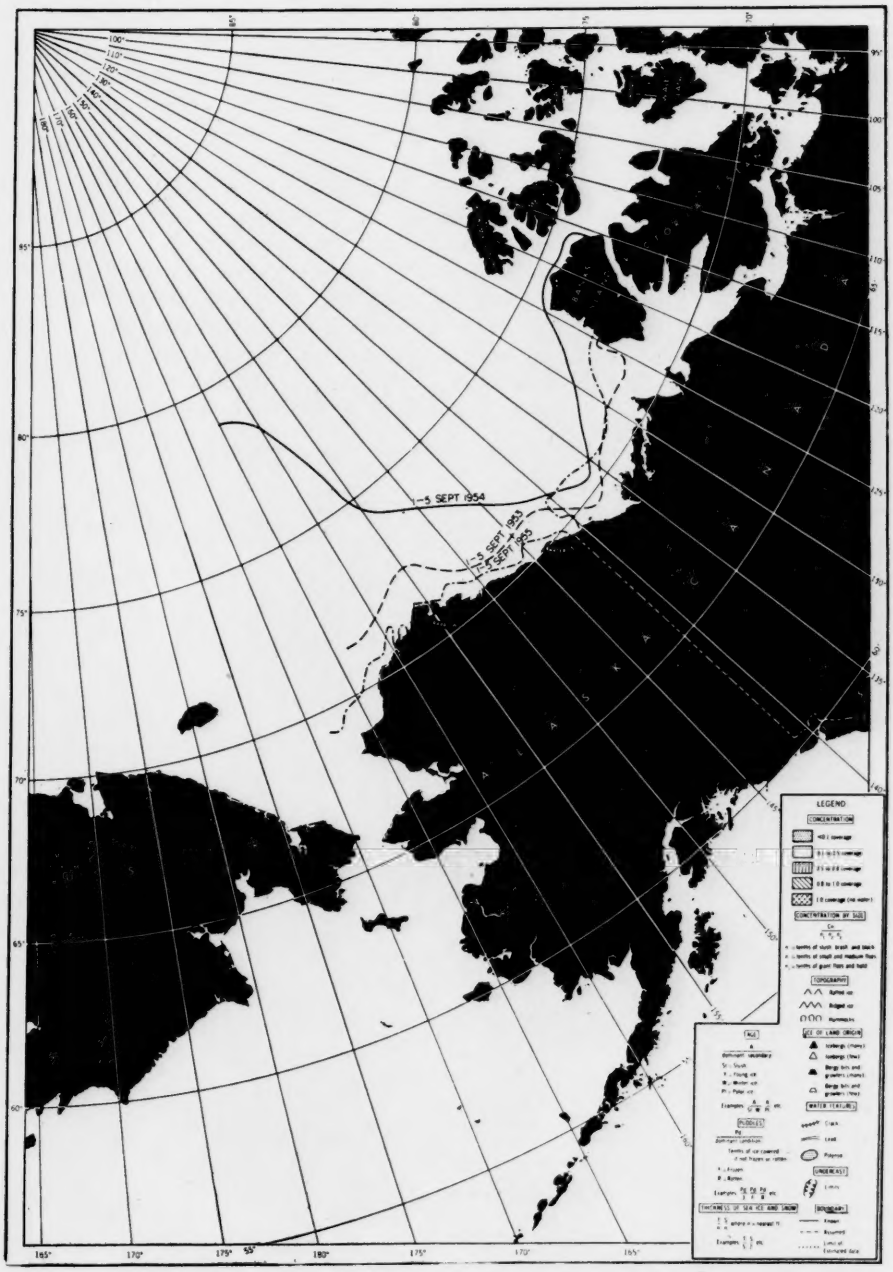


Fig. 15. Boundary of the Polar Pack off northern Alaska in early September of 1953, 1954 and 1955

Banks Island also was of a magnitude sufficient to affect markedly surface operations.

For most of the oceanographic circulatory systems of the Arctic there are still insufficient ice data to determine whether annual variations are the result of fluctuations in the total amount of ice present, in concentration of the ice, in a shift of the mass centroid of the ice in a given basin, or of a combination of all three. Comprehensive reconnaissance of a single oceanographic system for even a 3-year sample exists only for the Labrador Sea-Davis Strait-Baffin Bay complex. The rate of disappearance of ice in this area betwen March and September of 1953, 1954 and 1955 is shown in *Fig. 16.* It will be noted that about 400,000 square miles of ice existed each March, 1953 being the year of least ice, 1954 the year of most ice and 1955 an intermediate year. On the route to Thule, Greenland, however, 1955 was by far the best year for unescorted shipping, this being possible by late June, whereas even icebreaker escorted shipping could not reach Thule until early July during the other two years. The 1955 situation came about

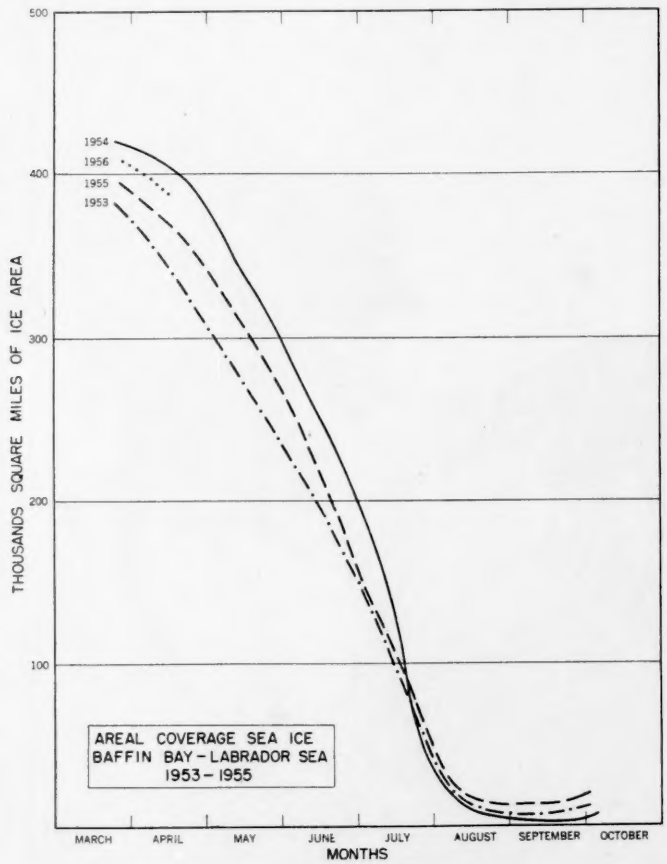


Fig. 16. March to October ice cover of Baffin Bay in 1953, 1954 and 1955

when the mass centroid of the Baffin Bay pack was shifted sufficiently westward during spring months so as to open prematurely the "Middle Passage"; it also created a worse-than-average year for shipping attempting to reach the eastern shore of Baffin Island, particularly the Home Bay sector.

In order to provide the requisite synoptic knowledge of ice conditions on which to base forecasts and ship scheduling, the U.S. Navy Hydrographic Office has sponsored an extensive ice-reporting system. Before this system could operate, it was necessary to develop an ice terminology giving more detail than then existing in the international meteorological reporting system. In 1952 a systematic glossary arranged in "check list" form was issued as Hydrographic Office Publication 609, *A Functional Glossary of Ice Terminology*. Next, a comprehensive shipboard and shore-station ice-reporting code were developed and tested (see *Tables II* and *III*). Shore reports are now made upon request by coastal stations of the U.S. Weather Bureau, Air Weather Service and the Canadian Meteorological Service. Shipboard observations are made at least once daily by a designated ship at each ice-infested location and radioed directly to the U.S. Navy Hydrographic Office. Aerial ice reconnaissance data are transmitted in one of two forms, numerical code or word message. The aerial ice reconnaissance code consists of seven five-digit groups (see *Table IV*) and has been proposed to the World Meteorological Organization as an addition to the present RECCO code. Currently, this code is used nationally whenever special ice observers are not aboard. Word messages are used by several dozen specially trained aerographers of the U.S. Navy Aerological Service, who have received a special six-week course in ice observing before being assigned for ice duty to the Fleet Weather Centrals at Argentia, Newfoundland, and Kodiak, Alaska. This method is described in H.O. Misc. 15603, *Aerial Ice Reconnaissance—Observational Techniques and Recording and Reporting Procedures* (second edition 1956). During the winter Navy reconnaissance aerographers assist oceanographers studying the manner in which various harbors freeze. To date, reports have been prepared on Terrington Basin, Labrador,[6] North Star Bugt[9] and Sondre Stromfjord,[12] Greenland.

Sea-ice forecasting can be broken down into five major fields of endeavor:

(*a*) Formation of sea-ice.
(*b*) Accretion of sea-ice.
(*c*) Distortion and metamorphism of sea-ice.
(*d*) Drift of sea-ice.
(*e*) Disintegration of sea-ice.

In the United States, the art of forecasting is still such that only the formation and accretion aspects can be said to be truly quantitative. Expanding upon work of ZUBOV[14] and DEFANT,[6] BROWN,[2] CALLAWAY,[3] CORTON,[5] and LEE and SIMPSON[8] of the Hydrographic Office have developed a highly useful forecasting system of "ice potential" based upon an initial knowledge of the temperature and salinity structure in the upper 100 m or so each autumn. The basic field data are obtained by Hydrographic Office oceanographers aboard icebreakers after the reversal of the heat budget at the close of the shipping season. Typical station coverage for the eastern Arctic is shown in *Fig. 17*. The oceanographic station data are then processed at the

Hydrographic Office by high-speed electronic computers to indicate the amount of heat that must be lost in order to form ice of any thickness. By utilizing air temperatures reported from nearby shore stations during the winter it is not only possible to calculate the position of the ice boundary rather closely but also actual thickness of the pack-ice to within several inches (see *Fig. 18*). A similar technique works for the freeze-up of protected

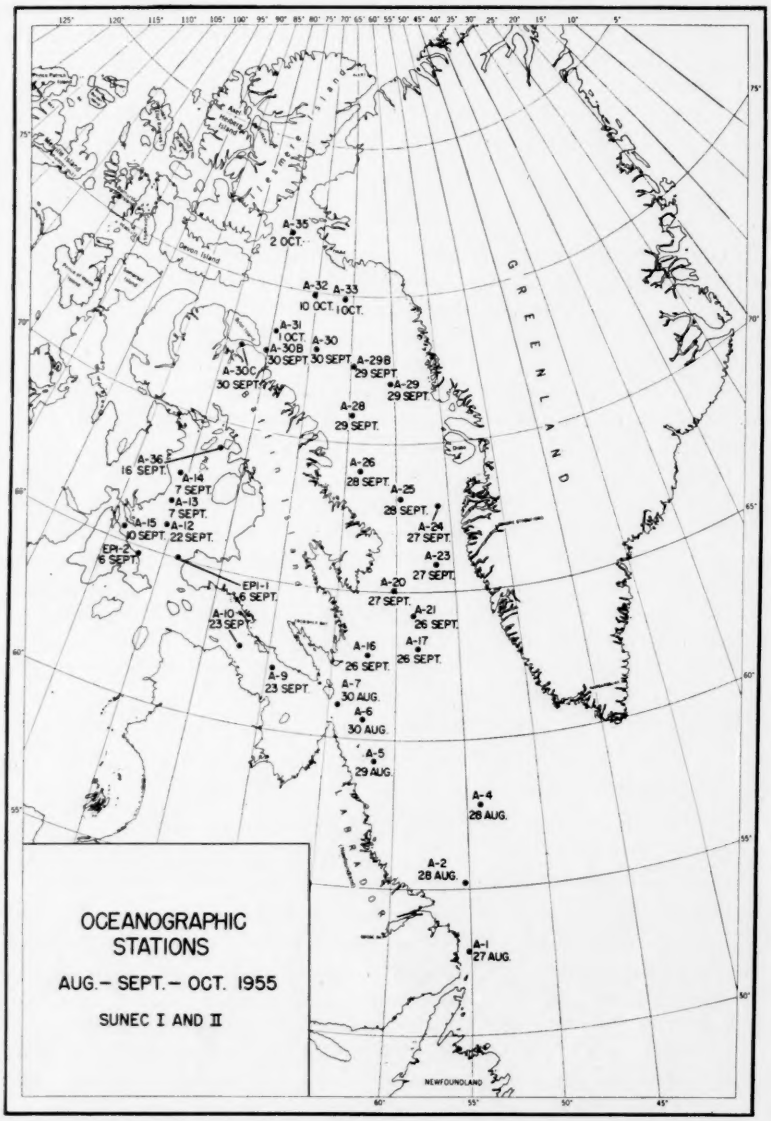


Fig. 17. Oceanographic stations occupied in 1955 for determining ice potential values in the eastern American Arctic

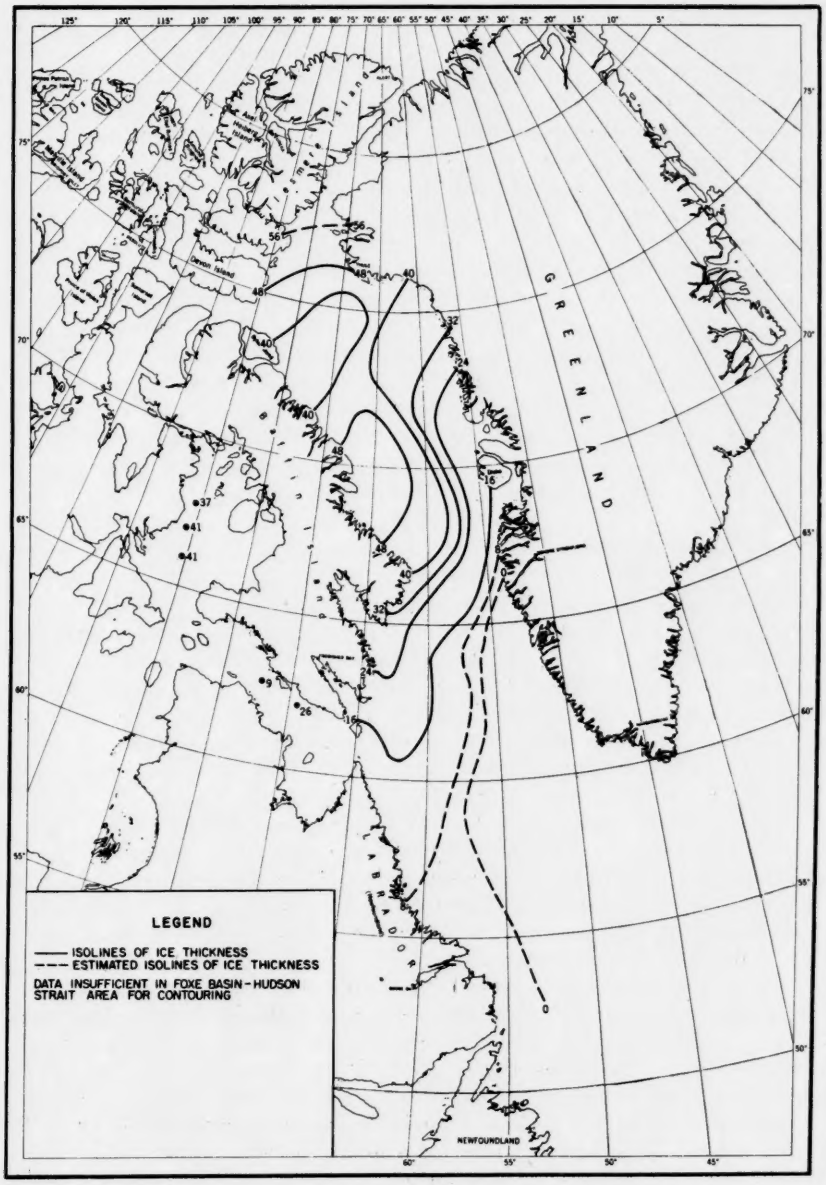


Fig. 18. Undisturbed ice thickness in inches as of 1 March 1956 calculated from ice potential values

Suppl.
Part.I
1956

coastal points. Seaplanes are substituted for ships in order to obtain rapidly the necessary oceanographic data on which to base freeze-up forecasts issued about 10 October giving dates of initial ice formation, and ice of 12-in. and 30-in. thicknesses along the Labrador coast.

Sixty-day forecasts of this type have been accurate to within a day and are of value in planning ship exit times as well as ski-plane operations. Accretion forecasts are primarily of the analog type. Once an ice cover is formed, and provided snow cover is negligible, ice thickness increases at a rate proportional to the number of degree-days below 32°F for fresh water or below 29°F for sea water. Corrections are applied for thickness and density of snow cover. Seasonal curves of degree-days versus ice thickness have now been acquired for a number of coastal points and provide a most appropriate forecasting tool for taking into account the idiosyncracies of each site.

Forecasting the distortion of sea-ice, such as hummocking, rafting and fracturing, and its metamorphism, particularly the vertical salinity profile and crystal structure, can only be considered in a formative stage. At the moment, the art is primarily qualitative and based on storm intensities and tracks and the age of ice. Better techniques will depend in part upon the knowledge of ice physics being developed by laboratories such as the U.S. Navy Electronics Laboratory and the U.S. Army Corps of Engineers' Snow, Ice and Permafrost Research Establishment. Forecasting ice drift and the associated fields of convergence/divergence from surface winds goes as far back as Nansen. Probably the most promising theoretical work now available is that of the Russian, Shuleikin, and that of the group at the Low Temperature Laboratory of Hokkaido University. Both approaches, however, do not give adequate correction for boundary conditions created by coastlines and shallow sea bottoms, although both the theory and empirical relations take into account variations in ice concentration and surface roughness. For example, in observing the polar pack off Alaska, it has been found that the pack's margin does not move at a rate exceeding six miles a day, or at less than half the rate normally associated with the drift of an isolated ice floe in an open sea during severe storms.

The importance of increased knowledge as to how ice drifts in a wind field can be illustrated by the type of forecasting required for supply operations attempted in April 1953 at Resolution Island. In this area the stream of pack-ice moving south out of Davis Strait is normally about 40 miles wide and has a concentration of 7/10 to 9/10. During this particular April, however, the mean 30-day map of the 700-mb surface showed a deviation of 220 ft below normal in the Great Lakes Region, while over the northern Baffin Bay area, the 700-mb surface was 610 ft above normal.[1] This condition resulted in a storm track to the south of the area and gave rise to consistently onshore winds during April. It also happened that the winter of 1952–53 was one of the warmest on record for the American Arctic, so that ice in the southern portion of the Baffin Bay pack broke up more readily under storm action and was advected south in quantities much greater than usual. The combination of this greater advection rate and the persistent onshore winds gave rise to the fluctuations in the boundary shown in *Fig. 19*, where it is noted that by 28 April 1953 the ice belt was constricted into a zone but 7 miles in width. Although this winter ice was well broken and

largely composed of block and small to medium floes, the concentration was so high that even a *Wind*-class icebreaker was severely handicapped in penetrating the band.

An adequate knowledge of disintegration processes is only now becoming available. In a regional sense, one of the most intriguing results is that there are probably six major centers in the American Arctic where a thick cover of sea-ice is lacking during the winter.[11] Their location is shown in *Fig. 20.* The most famous is the North Open Water off Thule, Greenland, which has an area of about 1600 square miles in March; here the water stays either open or, at best, has a foot-thick cover of young grey ice. Schule suggests that in this situation, as in the Prince Regent Polynya and the Igloolik Lead,

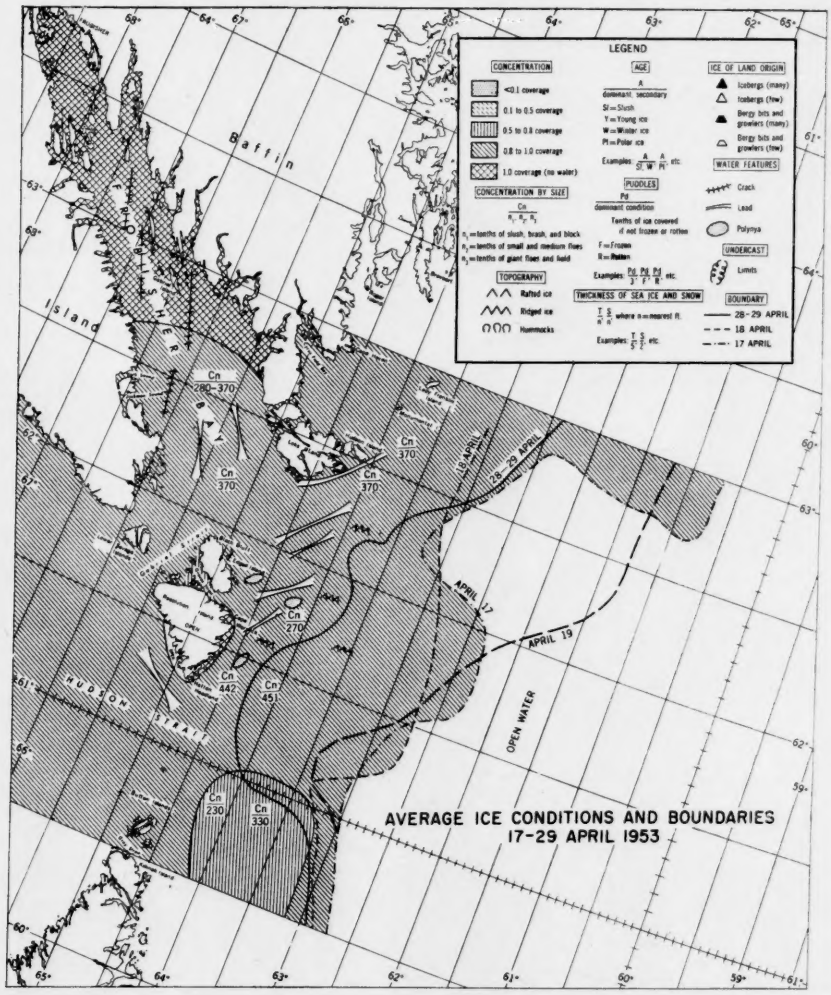


Fig. 19. Observed boundary of pack-ice off Resolution Island, Northwest Territories, 17–30 April 1953

Suppl.
Part. I
1956

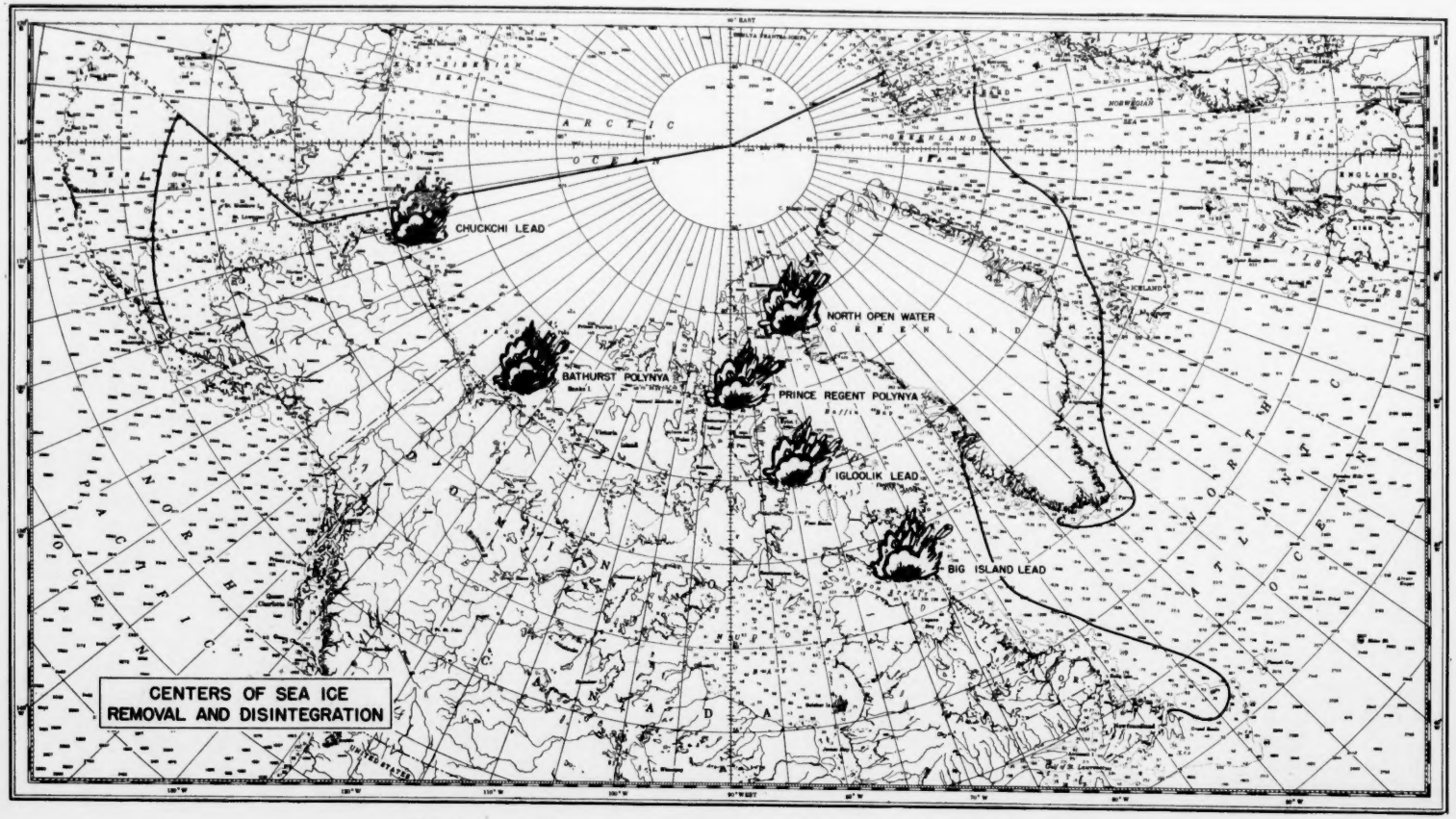


Fig. 20. Important centers of ice disintegration in the American Arctic (after Schule)

Suppl
Part
1956

strong northerly winds jam winter and polar ice into a relatively immovable mass just north of these areas, as well as drive southward any new ice formed in the area itself. The Big Island Lead is likewise probably related to strong offshore winds. The Bathurst Polynya is definitely associated with a bathymetric ridge, strong tidal currents and perhaps a divergent current. In the cases mentioned so far, upwelling is also suggested but difficult to prove until oceanographic data are obtained during the winter months. The last of these centers, the Chuckchi Lead, is thought to be primarily an expression of the flow of warm water north eastwards from the Bering Strait. One suggestion of this effect is that in March and April 1954 surface water temperatures ran as much as 5°F above average along the Aleutian Islands, and subsequently in May 1954 ice navigation was much easier than usual off northern Alaska.[10] The U.S. Navy Electronics Laboratory now maintains a field station at Cape Prince of Wales on the Bering Strait, and the

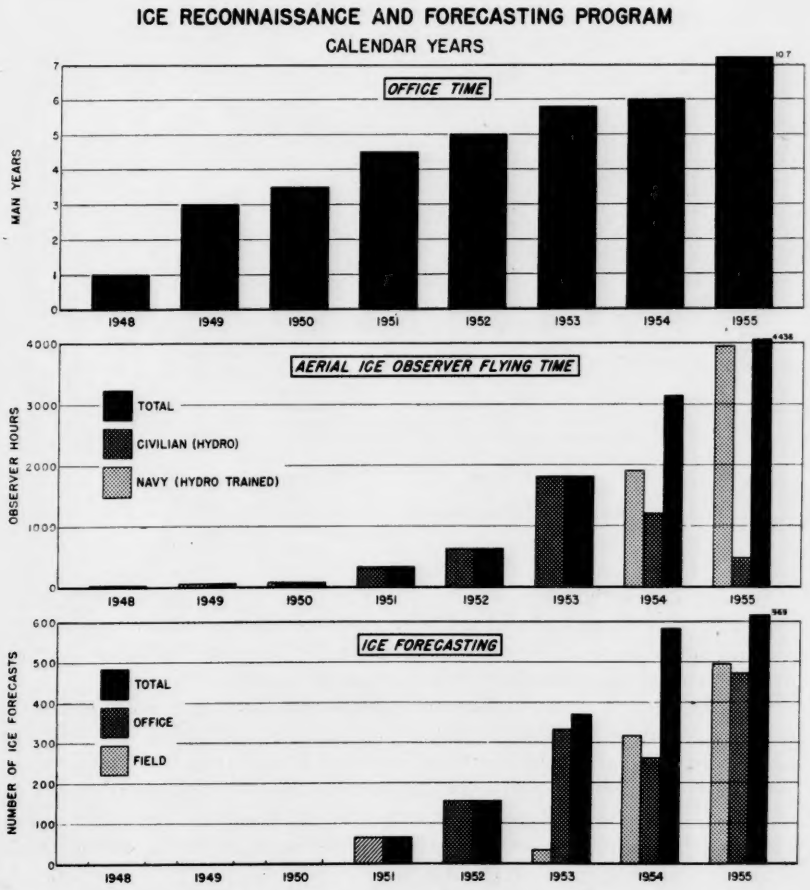


Fig. 21. Histograms portraying the U.S. Navy's increasing effort in observing and forecasting sea ice, 1948–55

temperature and current data obtained here should soon prove or disprove the correlation just proposed.

In the present state of ice forecasting, outlooks of greater than 30 days' duration are based on existing conditions and the recent climatology and oceanography of the area. Thirty-, 5 and 2-day forecasts become progressively more dependent on weather forecasts of the same duration. Thirty-day and 5-day weather forecasts are obtained directly from the U.S. Weather Bureau's Extended Forecast Section located adjacent to the Hydrographic Office. The 48-hr weather forecasts used in Washington, D.C., are a composite of those issued by the adjacent National Weather Analysis Center and the Joint Numerical Weather Prediction Unit, and that

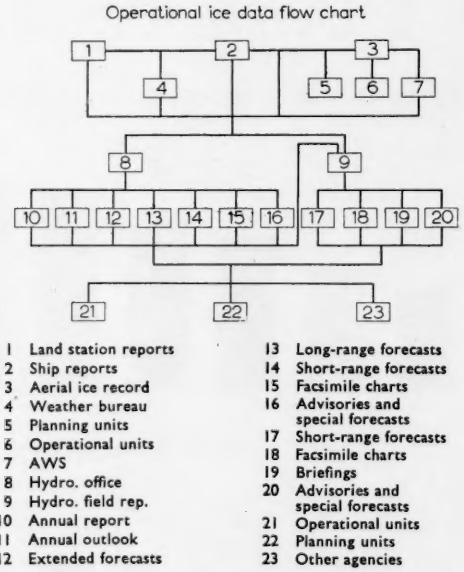


Fig. 22. Diagram of channels used to exchange ice observations and forecasts during an Arctic shipping operation

of the oceanographic forecaster himself. Forty-eight-hour ice forecasts, however, are more customarily issued by field forecasters stationed either aboard a command ship or at a weather station of the Navy or Air Force. The ice forecaster then uses weather forecasts available locally.

In order to provide the comprehensive ice reconnaissance and forecasting service so far described, the U.S. Navy had steadily increased since 1948 the amount of office and field effort and received in return a steadily increasing number of ice messages and forecasts (see *Fig. 21*). The present service is an extremely complex one based, as in meteorology, upon rapid communication circuits for transmitting word messages, numerical messages, and facsimile synoptic and prognostic charts. Arrangements for supporting an actual field operation are diagrammed in *Fig. 22.*

Although the long-range ice forecast program is but three seasons old,

verifications to date indicate that a skill factor is present. For example, during mid-August 1955 in the Baffin Bay-Labrador Sea-Foxe Basin area, there were about 385,000 square miles of ice. The deviations from the actual conditions observed by an ice atlas,[13] the 120-day ice outlook, and 30-day ice outlook for the actual conditions observed are shown in *Figs. 23, 24* and *25*, respectively. In terms of areal deviation, the ice atlas incorrectly blocked in 85,000 square miles, the 120-day outlook 48,000 square miles and the

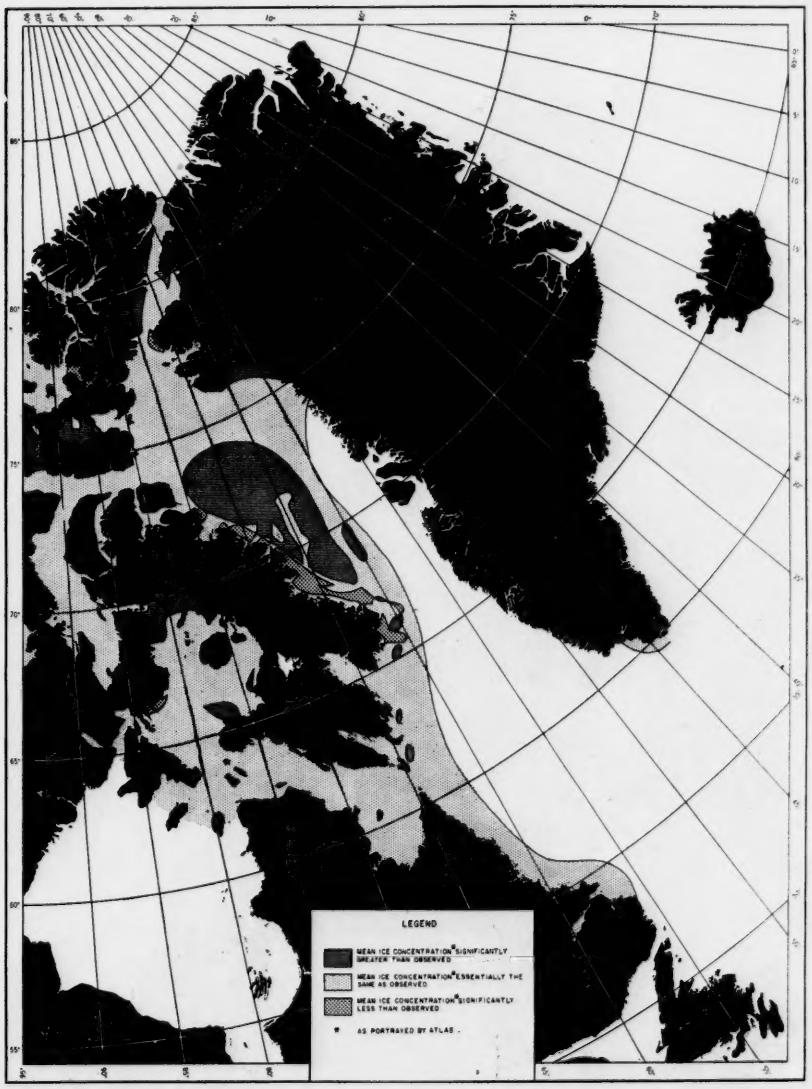

LEGEND
MEAN ICE CONCENTRATION SIGNIFICANTLY GREATER THAN OBSERVED
MEAN ICE CONCENTRATION ESSENTIALLY THE SAME AS OBSERVED
MEAN ICE CONCENTRATION SIGNIFICANTLY LESS THAN OBSERVED
AS PORTRAYED BY ATLAS

Fig. 23. Comparison between mid-August 1955 ice conditions with those portrayed in an ice atlas (H.O. Pub. 550)

30-day forecast 30,000 square miles. Results of another type of verification for the same area are shown in *Fig. 26.* In this situation a random sample of all the ice forecasts for the area has been compared against existing conditions; "significant deviation" is defined as being off by two or more categories of ice concentration. The results indicate not too great an improvement in the shorter range forecasts over the longer range outlooks; however, it is noteworthy that the former type of forecasts does not go as wildly "off target" as the latter type.

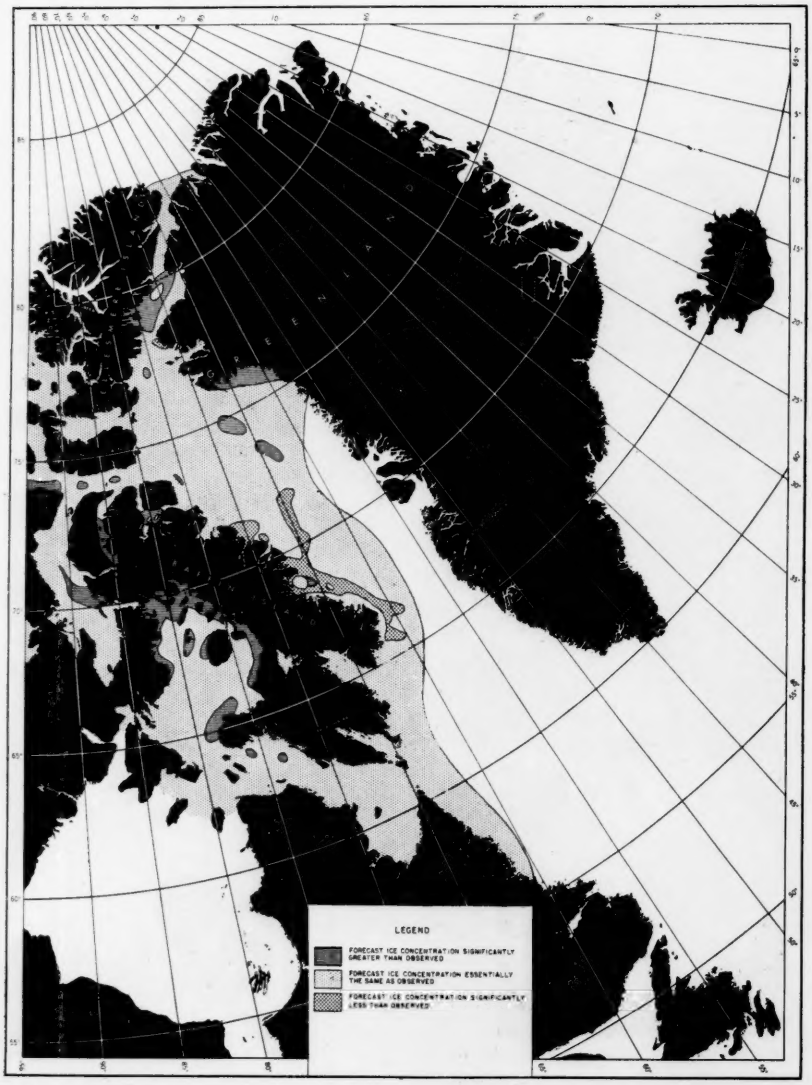

LEGEND
FORECAST ICE CONCENTRATION SIGNIFICANTLY GREATER THAN OBSERVED
FORECAST ICE CONCENTRATION ESSENTIALLY THE SAME AS OBSERVED
FORECAST ICE CONCENTRATION SIGNIFICANTLY LESS THAN OBSERVED

Fig. 24. Comparison between mid-August 1955 ice conditions and those portrayed in a 120-day forecast

Just as in other fields of scientific endeavor that have undergone rapid expansion, the sheer press of events has kept the pure research effort at the minimum and the operational aspect at the maximum. Accordingly, the visual approach to ice observation has dominated to date, but work on tracking individual ice floes has already begun. The most promising step in this direction is the use of radio-monitored weather buoys developed for the U.S. Navy Bureau of Aeronautics by the Naval Research Laboratory and the Bureau of Standards. The first of these buoys, weighing about 400 lb,

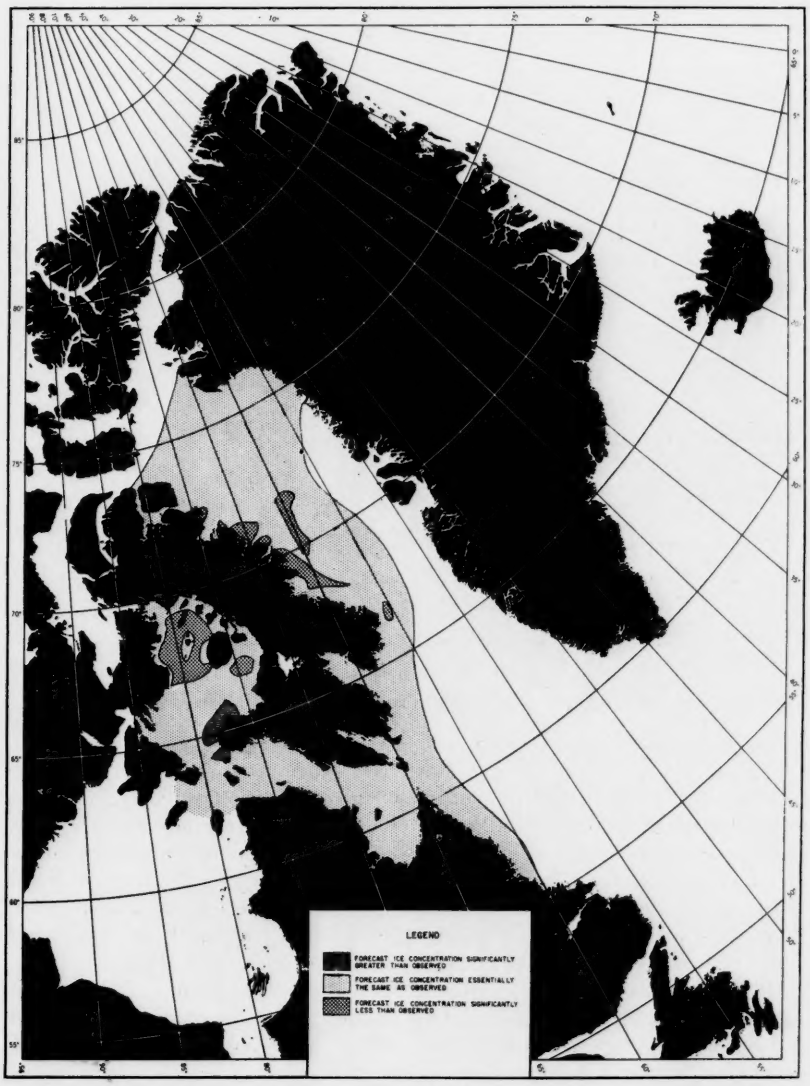

LEGEND
FORECAST ICE CONCENTRATION SIGNIFICANTLY GREATER THAN OBSERVED
FORECAST ICE CONCENTRATION ESSENTIALLY THE SAME AS OBSERVED
FORECAST ICE CONCENTRATION SIGNIFICANTLY LESS THAN OBSERVED

Fig. 25. Comparison between mid-August 1955 ice conditions and those portrayed in a 30-day forecast

Suppl.
Part.I
1956

was installed on an ice floe about 150 miles north of Point Barrow, Alaska, in April 1956, by using a light aircraft of the Alaskan Air Command. This buoy reports wind speed and direction, barometric pressure and air temperature every 6 hr and can be tracked by radio-direction finding equipment. More of these buoys have since been assigned the Hydrographic Office for use north of Alaska and in Baffin Bay. Another method of ice tracking now being attempted is that of following icebergs off the coast of Labrador with coastal radar of the Northeast Air Command. Time lapse photography with one exposure per minute has also been tried with encouraging results

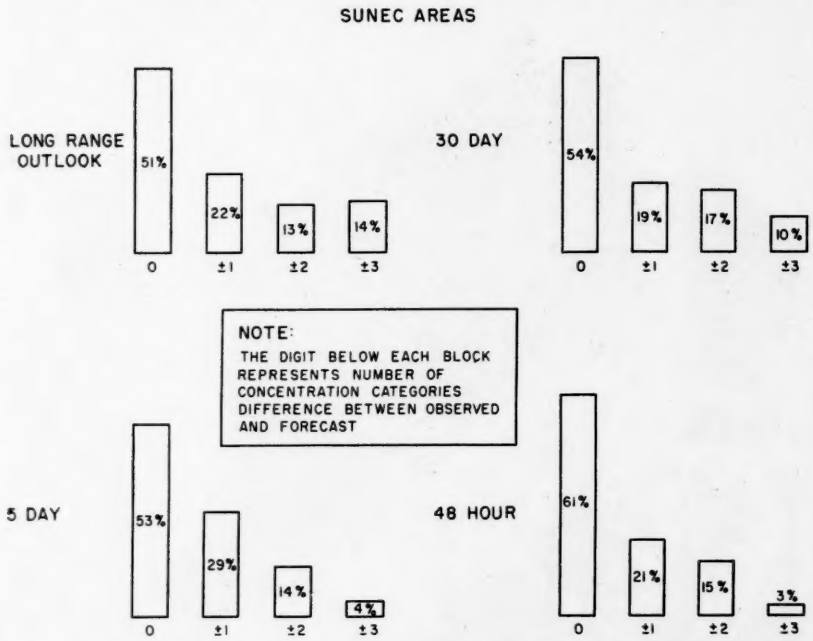


Fig. 26. Histograms showing percent verification of ice forecasts made for eastern American Arctic in 1955

for studying puddling and the consequent disintegration and flushing of ice from Hopedale Fjord, Labrador. Plans are also being made to utilize an airborne radiometer for detecting strong horizontal temperature gradients at the sea surface from ice reconnaissance aircraft.

In the field of ice physics, the most comprehensive program now under way is the joint Air Force-Navy-Army-Canadian Defense Research Board study concerning the strength of sea-ice with relation to landing heavy aircraft. This project has been conducted for the past two winters at Hopedale, Labrador. The oceanographic and micro-meteorological observations at the site are carried out by Hydrographic Office and Aerological Service personnel, while the Air Force Cambridge Research Center, assisted by consultants of the Corps of Engineers' Snow, Ice and Permafrost Research

Suppl.
Part. I
1956

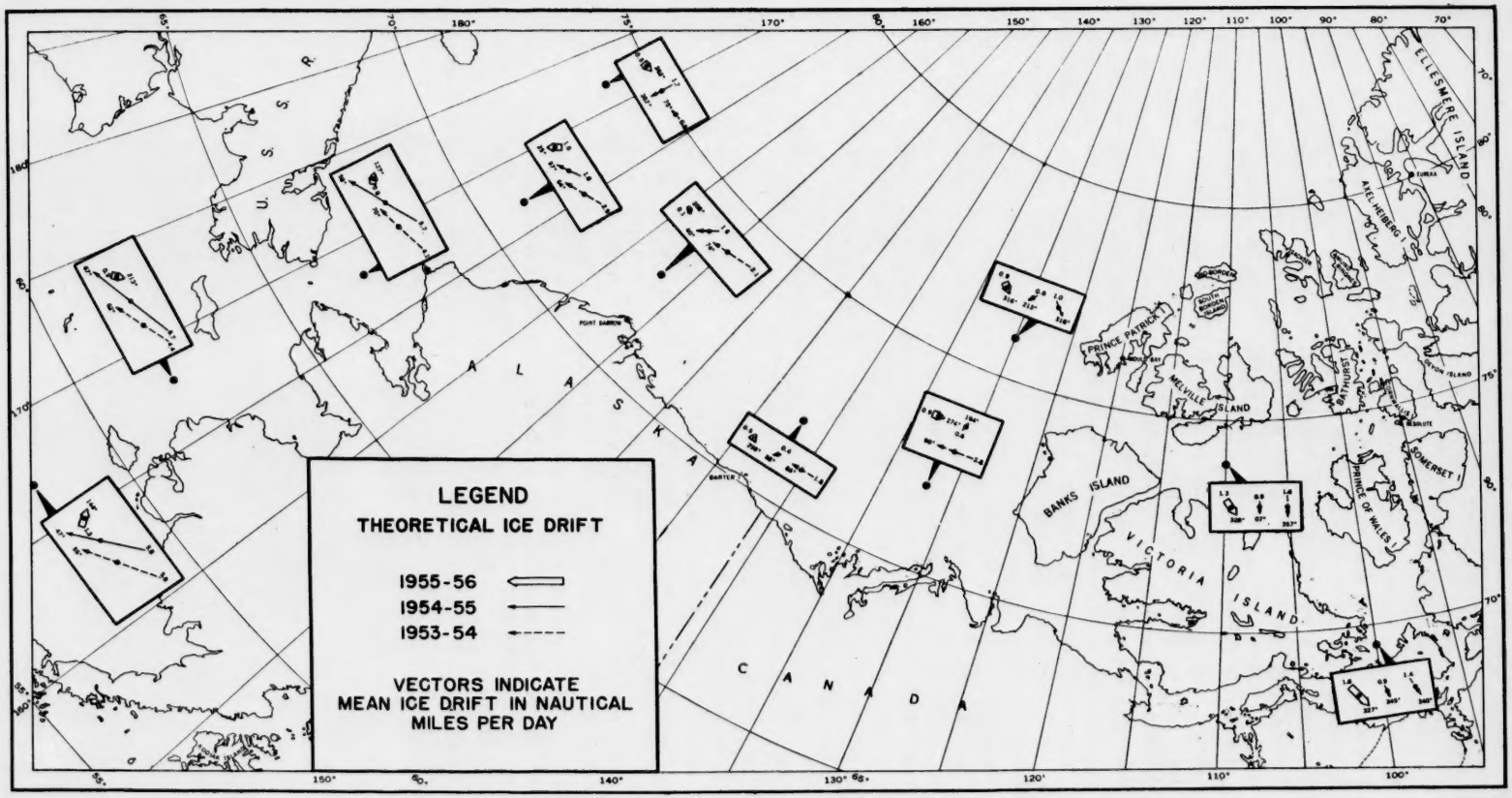


Fig. 27. Magnitude and direction of theoretical ice drift caused by wind between December and February, inclusive, for 1953, 1954 and 1955

Establishment, carries out engineering and seismic tests. The Defense Research Board of Canada supplies native aides, and the Northeast Air Command aircraft. Physicists of the U.S. Navy Electronics Laboratory have also studied ice physics from icebreakers in the Bering Sea, and an experimental tank has been constructed at their laboratory to permit freezing an ice sheet with a thickness comparable to that found in nature.

In the development of ice theory, the Hydrographic Office has experimented with computing mean monthly fields of surface wind divergence and convergence for Baffin Bay and for the Polar Basin. A by-product of this work is shown in *Fig. 27*, which shows the theoretical ice drift in Baffin Bay from December through February for each of the past three years. It appears that such computations could readily be made on the high-speed equipment of the Joint Numerical Weather Prediction Center as an addition to the programming required to compute the daily Northern Hemisphere 1000-mb prognostic pressure fields. Work has also been started on the ice budget of the Polar Basin[4] and is now being expanded by Teleki of the Arctic Institute of North America studying existing sets of aerial ice photographs on an Office of Naval Research Contract. In the meteorological field, the U.S. Weather Bureau has assigned a 30-day weather forecaster to specialize in the polar region. The feasibility of establishing some type of program for developing a 15-day mean pressure and temperature forecast is also under study.

In summary, it may be stated that there now exists a comprehensive ice surveillance and forecasting program for the American Arctic. Although this program is primarily still a responsibility of the U.S. Navy, a number of Canadian meteorologists and earth scientists have spent a considerable amount of time at the U.S. Navy Hydrographic Office and are becoming increasingly more active in providing ice information for this region. Each year the ice program becomes more quantitative in nature as research and instrumentation attempt to catch up to the operational requirements. From the point of view of a forecaster, ice forecasting is perhaps more satisfying than weather forecasting because most ice processes do not take place as rapidly as most meteorological processes, thereby giving time for considered thought. With most Arctic supply ships thin-skinned, shipmasters have rapidly developed a proper respect for sea-ice and urgently await the results of the scientists and engineers who seek to understand why "ice is where you find it".

References

[1] Bates, C. C., Kaminski, H. and Mooney, A. R. "Development of the U.S. Navy's Ice Forecasting Service, 1947–53, and its Geological Implications" *Trans. N.Y. Acad. Sci.* Vol. II, No. 16, pp. 162–174, 1954

[2] Brown, A. L. "An Analytical Method of Ice Potential Calculation" *U.S. Navy Hydrographic Office Tech. Rept.* No. 5, pp. 13, 1954

[3] Callaway, E. B. "An Analysis of Environmental Factors Affecting Ice Growth" *U.S. Navy Hydrographic Office Tech. Rept.* No. 7, pp. 31, 1954

[4] Corton, E. "The Ice Budget of the Arctic Pack and its Applications to Ice Forecasting" *U.S. Navy Hydrographic Office Tech. Rept.* No. 6, pp. 27, 1954

[5] ——. "Climatology of the Ice Potential as Applied to the Beaufort Sea and Adjacent Waters" *U.S. Navy Hydrographic Office Tech. Rept.* No. 30, pp. 16, 1955

[6] Defant, A. „Konvektion und Eisbereitschaft in polaren Schelfmeeren" *Geogr. Ann.* Arg. 31, pp. 25–35, 1949

[7] Lee, O. S. "Local Environmental Factors Affecting Ice Information in Terrington Basin, Labrador" *U.S. Navy Hydrographic Office Tech. Rept.* No. 24, 1956

[8] Lee, O. S. and Simpson, L. S. "A Practical Method of Predicting Sea Ice Formation and Growth" *U.S. Navy Hydrographic Office Tech. Rept.* No. 4, September 1954

[9] McGough, R. J. "Local Environmental Factors Affecting Ice Formation in North Star Bugt, Greenland" *U.S. Navy Hydrographic Office Tech. Rept.* No. 23, March 1956

[10] McQuain, R. H. "Ice Break-up, Bering Sea—1954" (Unpublished office report.) *U.S. Navy Hydrographic Office*

[11] Schule, J. "Comparison of Sea Ice Conditions in the American Arctic" (abstract) *Trans. Amer. geophys. Un.* Vol. 37, No. 3, 1956

[12] Tapager, J. R. D. "Local Environmental Factors Affecting Ice Formation in Sondre Stromfjord, Greenland" *U.S. Navy Hydrographic Office Tech. Rept.* No. 22, November 1955

[13] U.S. Navy. "Ice Atlas of the Northern Hemisphere" *Hydrographic Office Pub.* No. 550, 1945

[14] Zubov, N. N. "Morskie vody i l'dy. (Marine Water and Ice.)" Moskva *Gidrometeorologicheskoe Izdatel'stvo*, pp. 451, 1938

Suppl.
Part. I
1956

H.O. Misc. 15584
(Revised 1-55)

U. S. NAVY HYDROGRAPHIC OFFICE

SHIP ICE LOG

TABLE 1. Concentration: total and by size

No sea ice	0
1/10	1
2/10	2
3/10	3
4/10	4
5/10	5
6/10 or 7/10	6
8/10 or 9/10	7
10/10	8
Not determined	9

TABLE 2. Relief and Ice Thickness

No sea ice or not determined	0
Less than 3 in.	1
3 to 6 in.	2
6 to 12 in.	3
12 to 18 in.	4
18 in. to 2 ft.	5
2 to 3 ft.	6
3 to 5 ft.	7
5 to 8 ft.	8
Greater than 8 ft.	9

TABLE 3. Topography

No sea ice	0
Slush or pancake	1
Flat ice	2
Few rafted floes	3
Few pressure ridges	4
Few hummocks	5
Many rafted floes	6
Many pressure ridges	7
Many hummocks	8
Screw ice	9

TABLE 4. Age or Color

No sea ice or not determined	0
Grease ice (dark grease film)	1
Slush or pancake (gray)	2
Ice crust (gray)	3
Young ice (transparent black or ash gray)	4
Winter ice (aqua in section, white from vertical)	5
Polar ice (blue, contours rounded)	6

TABLE 5. Pudding

No sea ice or not determined	0
Rotten, disintegrating ice	1
Puddles joined, extensive cracking	2
Puddles melted through	3
Puddles—7/10 to 9/10 ice area	4
Puddles—4/10 to 6/10 ice area	5
Puddles—1/10 to 3/10 ice area	6
Puddles—less than 1/10 ice area	7
Puddles frozen	8
Ice without puddles	9

TABLE 6. Snow cover

Trace or no snow	0
Less than 3 in.	1
3 to 6 in.	2
6 to 12 in.	3
12 to 18 in.	4
18 in. to 2 ft.	5
Greater than 2 ft.	6
2 to 3 ft.	7
Snow in drifts	8
Not determined	9

TABLE 7. Water features

No sea ice or not determined	0
Ice in belts or patches	1
Ice evenly distributed	2
Polynya(s) in combination with leads or cracks	3
Lead(s) in combination with cracks	4
Single polynya	5
Single lead	6
2 or more cracks	7
Single crack	8
No water features	9

TABLE 8. Orientation of water features

No distinct orientation	0
NE-SW	1
E-W	2
SE-NW	3
N-S	4
Parallels shore at East	5
Parallels shore at South	6
Parallels shore at West	7
Parallels shore at North	8
Not determined	9

TABLE 9. Fast ice

Shore not observed	0
Shore entirely clear	1
Ice blocking less than 1/2 shore	2
Ice blocking more than 1/2 shore but with openings	3
Ice blocking entire shore	4
Not determined	9

TABLE 10. Ice of land origin

No land ice	0
Less than 100 growlers	1
100 or more growlers	2
Less than 50 bergy bits	3
50 or more bergy bits	4
Less than 20 icebergs	5
20 or more icebergs	6
Belt(s) of icebergs, bergy bits, or growlers	7
Not determined	9

TABLE 11. Visibility and Dimension

Less than 50 yds.	0
50 to 200 yds.	1
200 yds. to 1/4 naut. mile	2
1/4 to 1/2 naut. mile	3
1/2 to 1 naut. mile	4
1 to 2 naut. miles	5
2 to 5 naut. miles	6
5 to 10 naut. miles	7
10 to 25 naut. miles	8
25 or more naut. miles	9

TABLE 12. Water sky and Iceblink

Features not present	0
Feature to the NE	1
Feature to the E	2
Feature to the SE	3
Feature to the S	4
Feature to the SW	5
Feature to the W	6
Feature to the NW	7
Feature to the N	8
Not determined	9

Instructions for use of ice log are bound at back of pad.

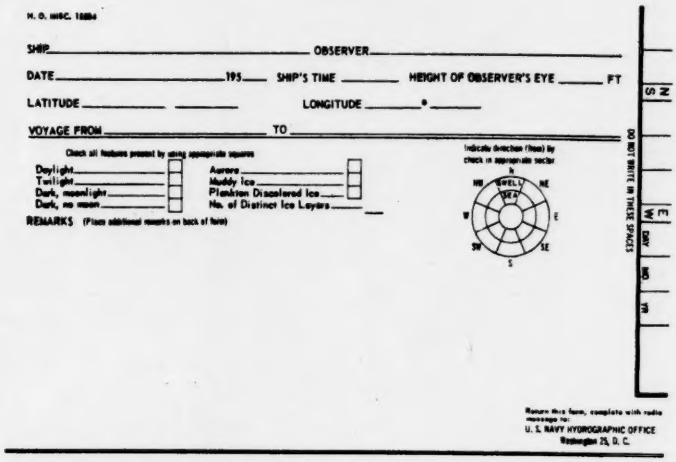
Table II. Supplemental Groups for Ice Boundaries and Water Features

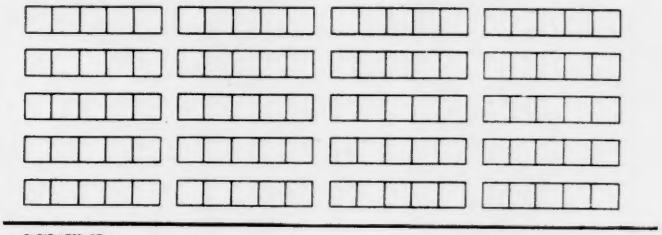
Group Indicators

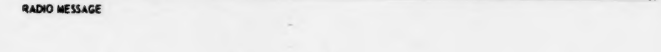
11000 Positions of drift ice boundary follow
22000 Positions of fast ice boundary follow
33000 Width and position of lead follow
44000 Dimensions and position of polynya follow

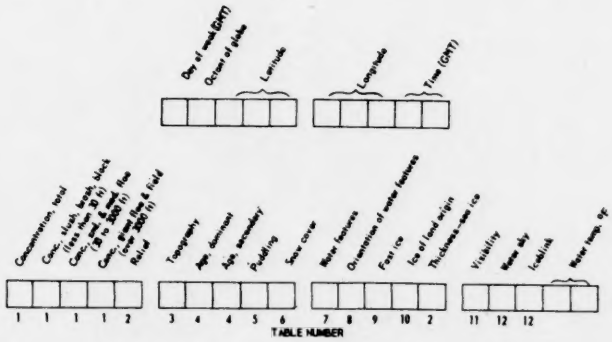
Order of Elements

$P L_a L_a L_a L_a$ $P L_o L_o L_o$	Points on boundary in degrees and minutes
$W L\ddot{a} L_a L_a L_a$ $W L_o L_o L_o L_o$	Width and position of lead in degrees and minutes
$D_w L_a L\ddot{a} L_a L_a$ $D_l L_o L_o L_o L_o$	Dimensions and position of polynya in degrees and minutes









Key of Elements

D_l Length of polynya (*Table 11*)
D_w Width of polynya (*Table 11*)
$L_aL_aL_aL_a$ Latitude in degrees and minutes
$L_oL_oL_oL_o$ Longitude in degrees and minutes
P Point on drift or fast ice boundary, coded 0
W Width of lead (*Table 11*)

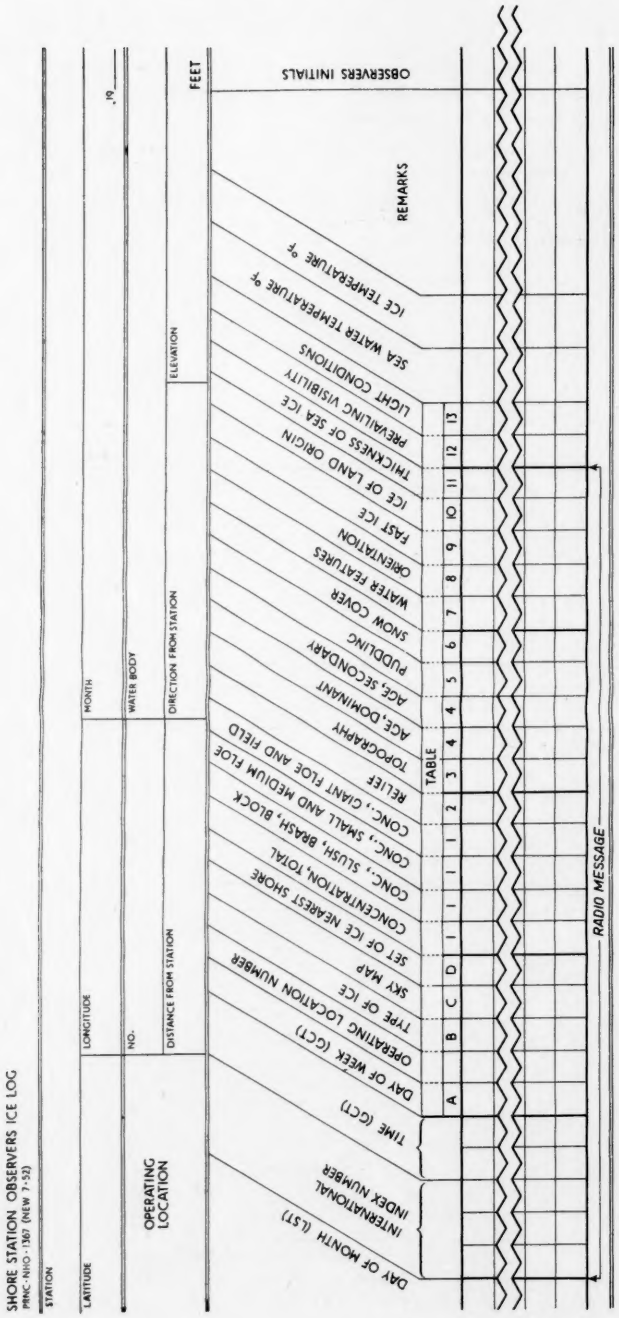

SHORE STATION OBSERVERS ICE LOG
PRNC-NHO-1367 (NEW 7-52)
STATION
LATITUDE
LONGITUDE
MONTH
19
NO.
WATER BODY
OPERATING LOCATION
DISTANCE FROM STATION
DIRECTION FROM STATION
ELEVATION
FEET
OBSERVERS INITIALS
REMARKS
ICE TEMPERATURE °F
SEA WATER TEMPERATURE °F
LIGHT CONDITIONS
PREVAILING VISIBILITY
THICKNESS OF SEA ICE
ICE OF LAND ORIGIN
FAST ICE
ORIENTATION
WATER FEATURES
SNOW COVER
PUDDLING
AGE, SECONDARY
AGE, DOMINANT
TOPOGRAPHY
RELIEF
CONC., GIANT FLOE AND FIELD
CONC., SMALL AND MEDIUM FLOE
CONC., SLUSH, BRASH, BLOCK
CONCENTRATION, TOTAL
SET OF ICE NEAREST SHORE
SKY MAP
TYPE OF ICE
OPERATING LOCATION NUMBER
DAY OF WEEK (GCT)
TIME (GCT)
INTERNATIONAL INDEX NUMBER
DAY OF MONTH (LST)
TABLE
A
B
C
D
1
1
1
1
2
3
4
4
5
6
7
8
9
10
11
12
13
RADIO MESSAGE

Table III. Tables for Shore Station Observers Ice Log

Table A. Day of week (GCT)

Sunday	1
Monday	2
Tuesday	3
Wednesday	4
Thursday	5
Friday	6
Saturday	7

Table B. Type of ice

No ice	0
Sea-ice	1
Bay ice	2
Fjord ice	3
Lake ice	4
River ice	5
Other	6

Table C. Sky map

Feature not present		0
Water sky to	East	1
	South	2
	West	3
	North	4
Snow or iceblink to	East	5
	South	6
	West	7
	North	8
Feature not determined		9

Table D. Set of ice nearest shore

No drift ice		0
Stations on open sea:		
Ice setting	East	1
	South	2
	West	3
	North	4
Stations on bays and fjords:		
Ice setting	Parallel to shore away from sea	5
	Parallel to shore toward sea	6
Either type station:		
All ice motionless		7
Set not determined		9

Table 1. Concentration: Total and by size

No ice	0
Open water—less than 1/10	1
Scattered ice—1/10 to 5/10	2
Broken ice—5/10 to 8/10	3
Close ice—8/10 to 10/10	4
Consolidated ice—10/10. No sea surface	5
Not determined	9

Table 2. Relief

No sea-ice	0
Less than 6 in.	1
6 in. to 2 ft	2
Greater than 2 ft	3
Not determined	9

Table 3. Topography

No sea-ice	0
Slush or ice cakes	1
Flat ice	2
Rafted ice	3
Ridged ice	4
Hummocks	5
Screw ice	6
Not determined	9

Table 4. Age

No sea-ice	0
Slush, pancake or ice crust	1
Young ice	2
Winter ice	3
Polar ice	4
Not determined	9

Table 5. Puddling

No puddles	0
Puddles frozen	1
Puddles—less than 1/10 ice area	2
Puddles—1/10 to 3/10 ice area	3
Puddles—greater than 3/10 ice area	4
Puddles—"melted through"	5
Puddles joined, extensive cracking	6
Rotten, disintegrating ice	7
Puddling not determined	9

Table 6. Snow cover

Trace or no snow	0
Less than 6 in.	1
6 to 12 in.	2
12 to 18 in.	3
18 to 24 in.	4
24 to 30 in.	5
Over 30 in.	6
Snow in drifts	7
Continuous snow cover	8
Snow not determined	9

Table 7. Water features

Open water or scattered ice	0
Broken ice	1
Two or more polynyas	2
Two or more leads	3
One polynya	4
One lead	5
Two or more cracks	6
One crack	7
No water features	8
Not determined	9

Table 8. Orientation

No distinct orientation	0
NE–SW	1
E–W	2
SE–NW	3
N–S	4
Parallels shore at East	5
Parallels shore at South	6
Parallels shore at West	7
Parallels shore at North	8
Not observed	9

Suppl.
Part. I
1956

Table 9. Fast ice

Shore not observed	0
Shore entirely clear	1
Ice blocking less than half shore	2
Ice blocking more than half shore but some openings	3
Ice blocking entire shore	4
Not determined	9

Table 10. Ice of land origin

No land ice	0
Less than 100 growlers	1
100 or more growlers	2
Less than 50 bergy bits	3
50 or more bergy bits	4
Less than 20 icebergs	5
20 or more icebergs	6
Not determined	9

Table 11. Thickness of sea ice

No sea-ice	0
Less than 4 in.	1
From 4 in. to 1 ft	2
From 1 ft to 3 ft	3
From 3 ft to 6 ft	4
From 6 ft to 8 ft	5
From 8 ft to 10 ft	6
From 10 ft to 12 ft	7
More than 12 ft	8
Not determined	9

Table 12. Prevailing visibility

Less than 50 yards	0
50–200 yards	1
200 yards–¼ nautical mile	2
¼–½ nautical mile	3
½–1 nautical mile	4
1–2 nautical miles	5
2–5 nautical miles	6
5–10 nautical miles	7
10–25 nautical miles	8
25 or more nautical miles	9
Not determined	X

Table 13. Light conditions

Daylight	1
Twilight	2
Night—moon and aurora	3
Night—moon	4
Night—aurora	5
Night—no moon or aurora	6

Return completed forms to, and order blank forms from:

U.S. NAVY HYDROGRAPHIC OFFICE
WASHINGTON 25, D.C.

Table IV. U.S. Navy Hydrographic Office Aerial Ice Reporting Code

3400.3 Groups $9A_TA_1A_2A_3$ TPW_fOS_i 1BBDD 2BBDD $311w_dw_d$ $411w_dw_d$ $511w_dw_d$

9	Indicator identifying sea-ice data
A_T	Total concentration of ice
A_1	Amount of slush, brash and blocks (less than 30 ft in size)
A_2	Amount of small and medium floes (30–3,000 ft in size)
A_3	Amount of giant floes and ice field (over 3,000 ft in size)
T	Topography
P	Puddles
W_f	Water features
O	Orientation of water features
S_i	Amount of fast ice
1	Indicator for drift-ice boundary
2	Indicator for fast-ice boundary
BB	Bearing to the nearest boundary in tens of degrees from the coordinates indicated in the fore part of the RECCO report
DD	Distance to the nearest boundary in nautical miles from the coordinates indicated in the fore part of the RECCO report
3	Indicator for a lead
4	Indicator for a polynya
5	Indicator for an ice belt
11	Length of major axis in nautical miles of the water feature (W_f)
w_dw_d	Width of the water feature (W_f) in nautical miles

Suppl.
Part. I
1956

METEOROLOGICAL CONDITIONS AND THE ASSOCIATED SEA ICE DISTRIBUTION IN THE CHUCKCHI SEA DURING THE SUMMER OF 1955

JAMES W. WINCHESTER and CHARLES C. BATES*
U.S. Navy Hydrographic Office, Washington, D.C.

A combination of circumstances, some of which are not entirely understood, combined to keep the polar pack-ice from receding off the northern coast of Alaska during the summer of 1955. As a result, the worst ice year in at least the past 11 years occurred, and over forty ships of the U.S. Navy's Task Force 5 had their exit blocked for 2 weeks at what was normally the height of the shipping season.

Although the 30-day weather forecast prepared in early August 1955 indicated normal surface warming and frequent offshore winds in the Chuckchi Sea area, by 14 August an unusually deep polar vortex was forming which steered the frequent Siberian storms either into the western Canadian Archipelago or the Polar Basin. Such a mean storm track gave a steady onshore flow of cold moist air which markedly slowed disintegration of the pack-ice, closed the Chuckchi Lead, gave rise to appreciable amounts of snow as early as 20 August at Point Barrow, and caused sporadic re-freezing of the leads by mid-August; however, the permanent reversal of heat budget did not occur until early October. It was not until this polar vortex began to break down about 5 September that storms did take a more southerly track across the Bering Sea and central Alaska. Once this more normal track was established, the stagnation of a 975 mb storm in the Bering Sea on 5–8 September created, at least in part, the offshore lead which permitted supply operations to again be conducted along the northern Alaskan coast. The behavior of the polar pack's margin under these anomalous meteorological conditions is described and sample verifications of ice forecasts given.

DURING the summer of 1955, Task Group 5.1 was required to deliver a large amount of supplies for construction of the Distant Early Warning (DEW) Line at sites between Point Barrow, Alaska, and Sheperd Bay, 1700 miles to the east in the central Canadian Archipelago. To accomplish this mission, over forty ships of the Task Group moved east of Point Barrow, chiefly during the first week of August 1955. By mid-August 1955 meteorological and ice conditions started to deteriorate along the 400-mile stretch of coast between Cape Lisburne, Alaska, and Herschel Island, Northwest Territories, and by 25 August 1955 shipping found it impossible to move past Point Barrow. Blocking of the return route lasted 2 weeks at the height of the shipping season, thereby jeopardizing the success of this supply mission. Although the operational aspects have been described by DENEBRINK[2] and BATES,[1] the meteorological conditions contributing to this unusually severe ice condition

* The opinions expressed in this paper are those of the authors as individuals and are not to be construed as necessarily reflecting the official views of the Navy Department.

have yet to be described in some detail. It is the purpose of this paper to provide such a description.

While all factors controlling the distribution of pack-ice in the Polar Basin are not completely understood, local phenomena listed below are known to play important roles in causing rapid disintegration of the polar pack-ice (i.e. polar ice with a concentration of 8/10 or greater) and in opening of the Chuckchi Lead to Point Barrow each summer:

(*a*) Storms moving over the Bering and Chuckchi Seas will either accelerate, decelerate, or even reverse the flow of comparatively warm Bering Sea water through Bering Strait. Current measurements made by the U.S. Navy Electronics Laboratory in Bering Strait during August 1949 indicate that short-period changes observed in the magnitude of both surface and bottom currents were related to wind stress over the two seas. Because of this, the net transport of water into the Chuckchi Sea may vary from a negative value to more than 50,000 m^3sec^{-1}. In July and August, surface-water temperatures of the Bering Sea range between 40° and 50°F; such water is therefore an extremely important heat source for ice disintegration when mixing occurs with the ice-infested waters off north-western Alaska.

(*b*) Storms creating dominant south and south-east winds over the Bering Sea usually produce offshore winds along the northern coast of Alaska as well; such winds create another heat source for ice disintegration, namely, air that has had a long trajectory over land warmed by nearly continuous daylight. Conversely, polar storms capable of decelerating the flow through Bering Strait normally move eastward and give cold, onshore winds that decrease ice disintegration.

(*c*) Surface stresses exerted on the polar pack by strong offshore winds move the pack's margin seaward; conversely, strong onshore winds move the margin landward.

The combination of these local effects is enhanced by the fact that all tend to operate together in phase. Because of the shortness of the shipping season, the location of any persistent storm track becomes of paramount importance. The problem is further complicated because identical ice conditions do not exist each spring when ice disintegration starts. Furthermore, little is known as to how the 2000-mile wide polar pack behaves as an entity from season to season, or how the pack behaves when several storms pass simultaneously over its surface.

Early ice reconnaissance conducted along northern Alaska in May 1955 indicated generally favorable conditions (see *Fig. 1*). The ice outlook issued by the Experimental Oceanographic Forecast Central of the U.S. Navy Hydrographic Office called for a slightly better shipping season than 1953, a season in which ice conditions were severe enough off north-eastern Alaska to cause a naval task unit to cut its planned survey schedule short by a month. During June 1955 mean air temperatures stayed normal or nearly so along northern Alaska (see *Fig. 2*). July 1955, however, saw monthly temperatures 2°F below normal at Point Barrow and at Barter Island, 300 miles to the east, even though there had been abnormal warming over Alaska between 18 and 26 July 1955. In fact Fairbanks, near 65°N latitude, recorded a maximum temperature of 93°F on 25 July 1955. Comparison of 1955's

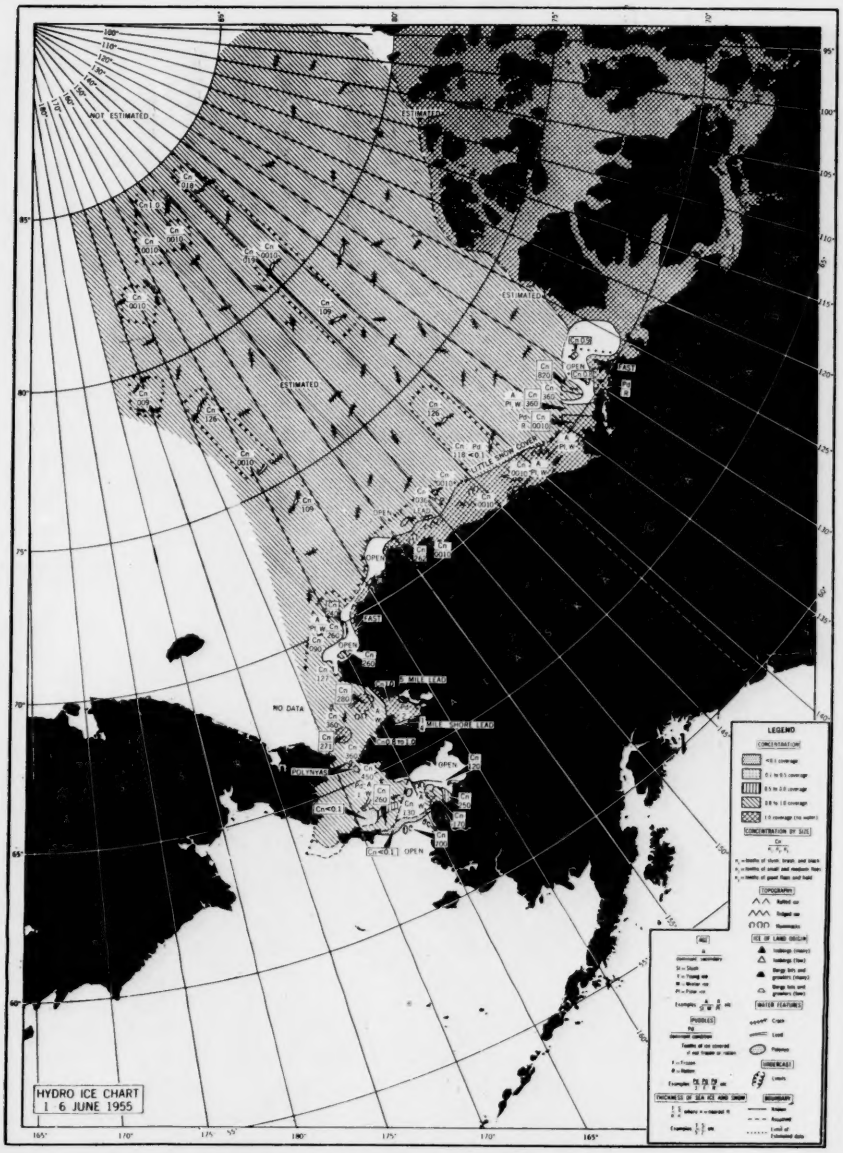


Fig. 1. Observed ice conditions 1–6 June 1955 in the Chuckchi and Beaufort Seas

mid-July boundary of the polar pack with that of 1953 and 1954 (see *Fig. 3*) indicated that in both 1953 and 1955 the Chuckchi Sea boundary lay rather close inshore as far south as Cape Lisburne, whereas in the extremely open year of 1954 the pack's boundary extended north-westward into the Chuckchi Sea from the vicinity of Wainwright. In the Point Barrow area proper, the

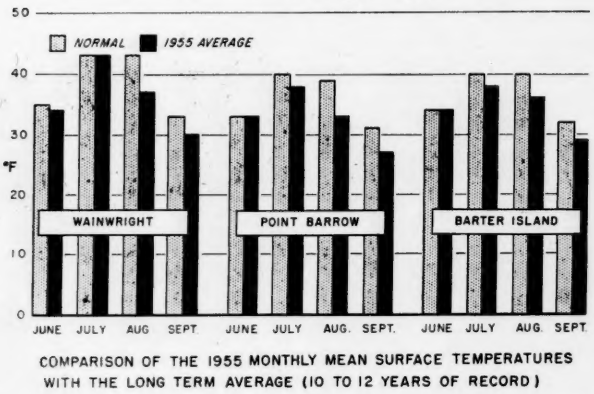


Fig. 2. Mean monthly surface air temperatures at north Alaskan coastal points during the summer of 1955

1955 boundary lay somewhat farther north than before; in the Beaufort Sea, 1953 and 1955 boundaries were similar, but again the 1954 boundary lay farther offshore between Barter and Banks Islands. On the basis of hindsight, future mid-July similarities in the boundary of the polar pack with those experienced in 1953 and 1955 suggest severe ice conditions in August and September unless such a trend is counteracted by a persistently favorable atmospheric pressure pattern.

The warm spell experienced during the last 2 weeks of July 1955 in Alaska caused improvement of ice conditions along northern Alaska, as shown by the comparison of early August polar pack boundaries in *Fig. 4* where the 1955 boundary is farther north off Alaska than in the other 2 years. As late as 8 August 1955 the 30-day weather forecast called for continued warming; consequently, the 30-day ice forecast prepared as of that date called for improving ice conditions and at least a normal ice year. Mean storm-track patterns for August show that the vast majority of North Pacific Ocean storms move first into the Bering Sea and then turn either into Bering Strait or move inland over Alaska just north of the Aleutian Islands. In the Polar Basin proper, storms often stagnate in the area of the 180° meridian and then drift slowly northward. However, between 1 and 15 August 1955 most of the North Pacific Ocean storms moved south of the Aleutian Islands into the Gulf of Alaska, thereby creating a deficiency of storms in the Bering Sea. Consequently, no storm centers moved northeastward along the northern coast of Alaska. On the other hand, the one major storm that did move eastward across the Polar Basin continued into the Canadian Archipelago, giving rise to cold, onshore winds that persisted after the storm's passage past Alaska. After several days of deliberation, the 30-day ice forecast was revised on 19 August 1955, in part, as follows:

> Special analysis mid-August ice conditions north Alaska coast. Ice recon conducted 16–17 August shows ice conditions more severe than during like periods 1953 1954. . . . Present state ice and predicted trend suggest slow clearing shore lead Cross Island to Brownlow Point X. Unless ice conditions rapidly improve this area during period 15–31 August early closing may be expected X.

This anomalous weather pattern was also appearing at the upper levels,

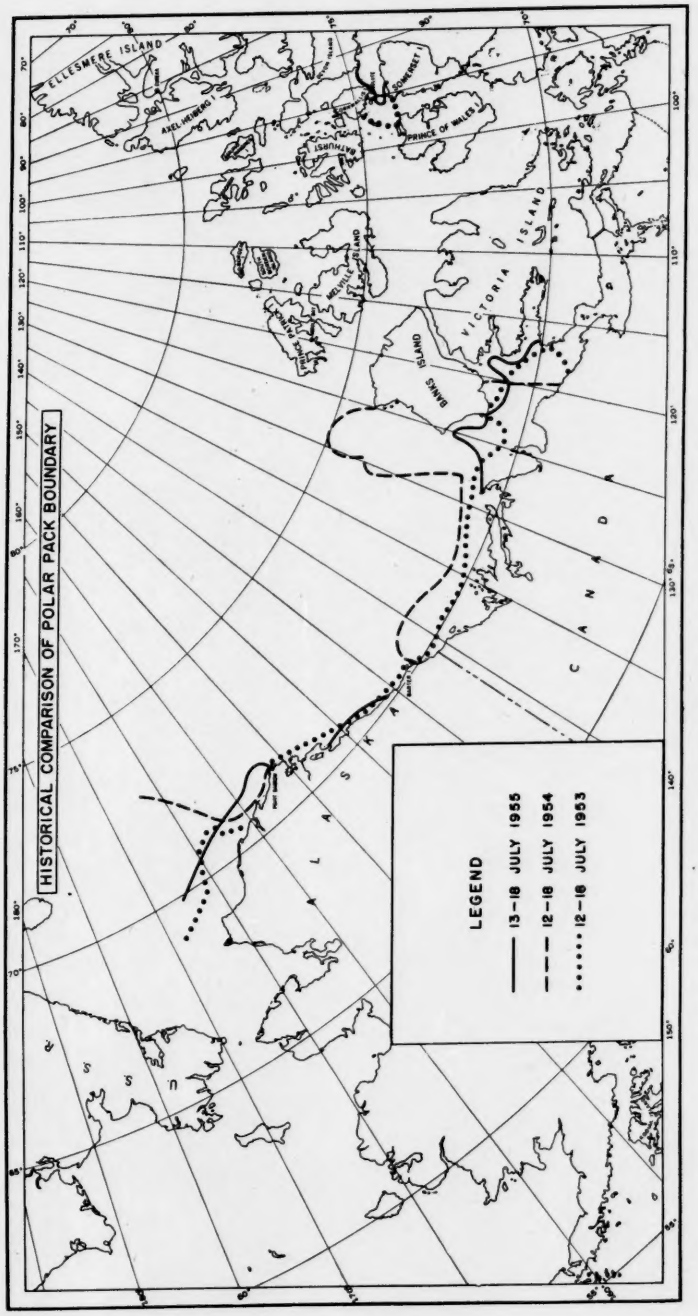


Fig. 3. Mid-July boundaries of the Polar Pack in 1953, 1954 and 1955 off northern Alaska

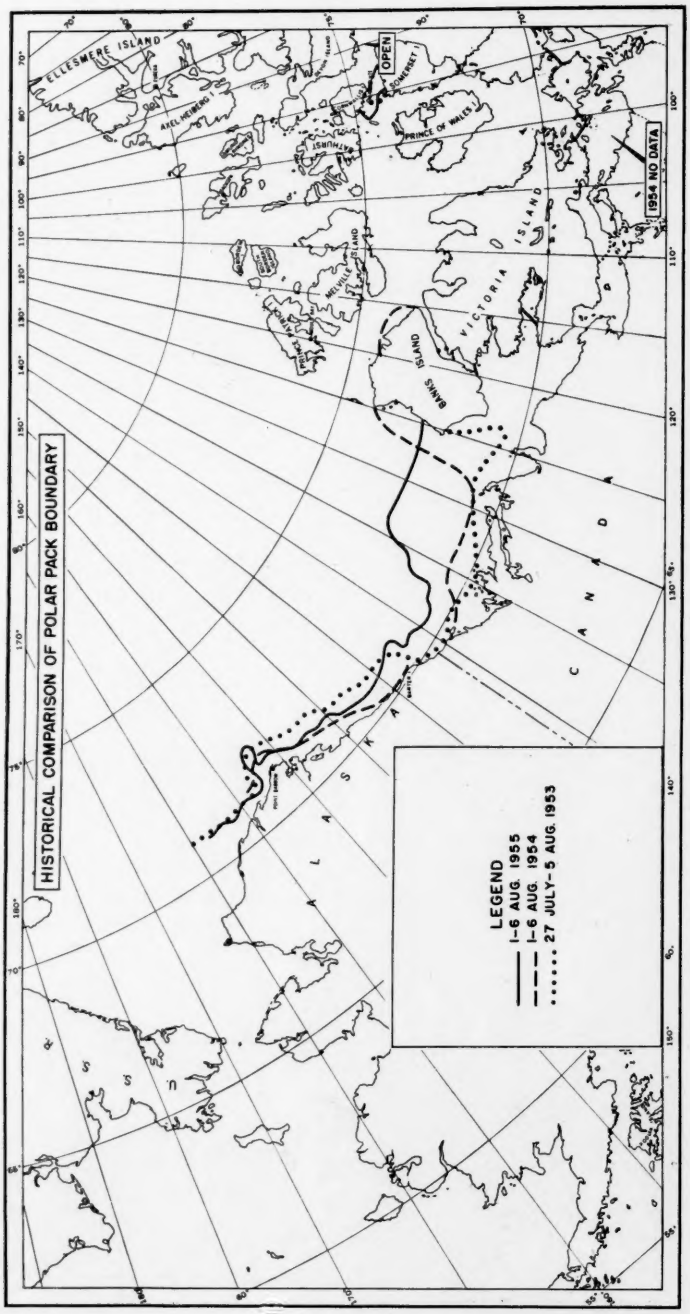


Fig. 4. Early August boundaries of the Polar Pack in 1953, 1954 and 1955 off northern Alaska

Suppl.
Part. I
1956

(Facing page 328)

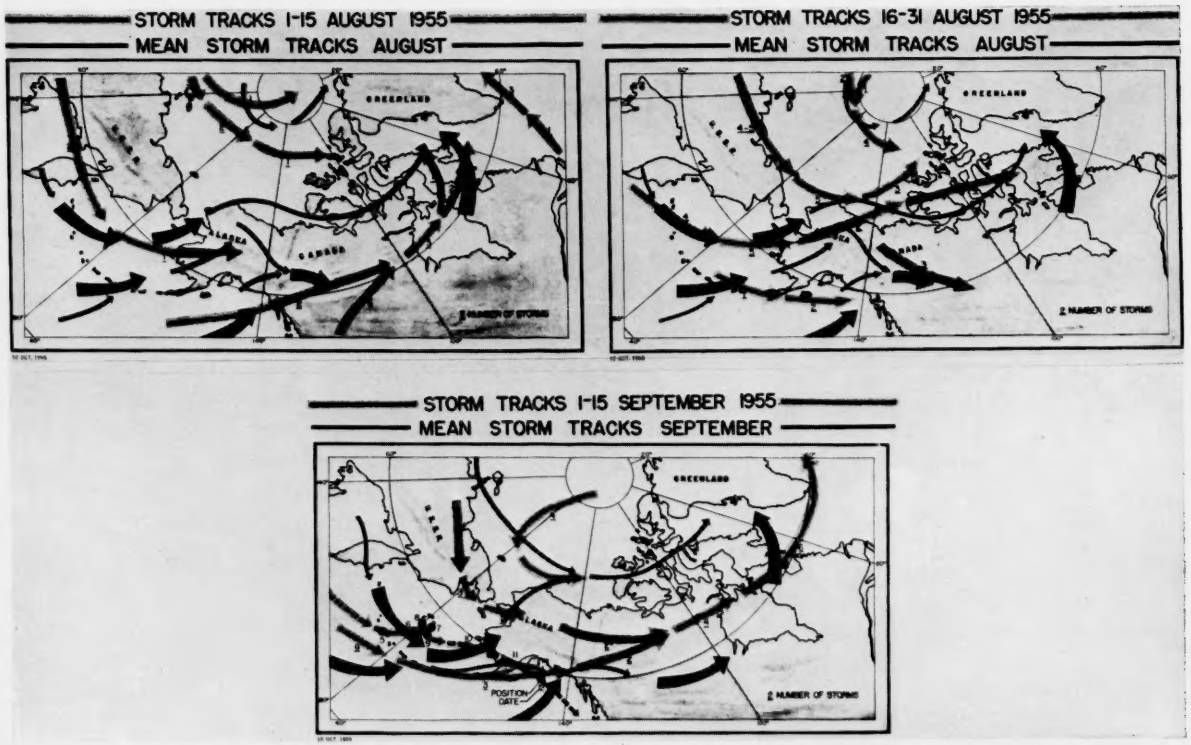


Fig. 6. Mean storm tracks for August (open arrows) and mean observed storm tracks for 15–30 August 1955 (solid arrows)

Sup
Par
195

and the mean 700-mb chart for 17–21 August 1955 showed a notably intensified polar vortex centered about 500 miles north of Alaska, the 700-mb surface level being 690 feet below normal (see *Fig. 5*).

Storm paths for the last two weeks of August 1955 superimposed on the mean August storm tracks are shown in *Fig. 6*. The two storms that dominated

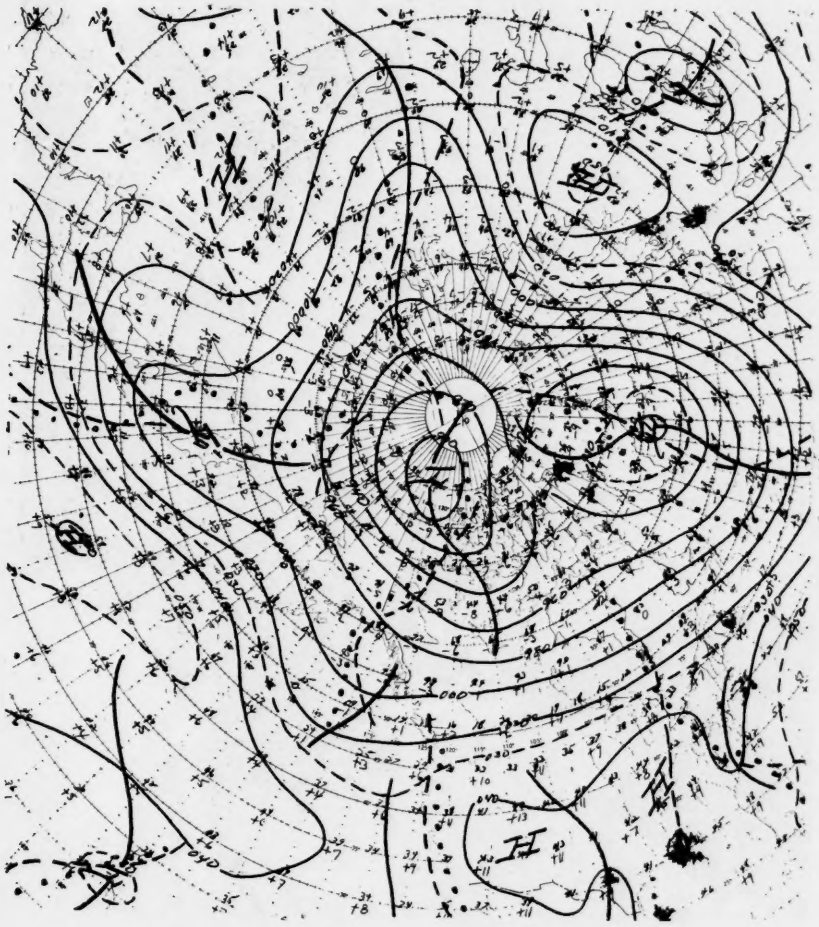


Fig. 5. Mean 5-day 700-mb level chart for 17–21 August 1955. (Courtesy U.S. Weather Bureau)

the weather picture during this time moved eastward through the Polar Basin, giving strong onshore winds along the entire northern Alaskan coast. Snow flurries became common in the third week of August, and new ice was reported forming off Point Barrow on 23 August 1955. The margin of the polar pack moved close inshore from Wainwright to Herschel Island, and during the passage of the more southerly of the two storms on 25 August

Suppl.
Part.I
1956

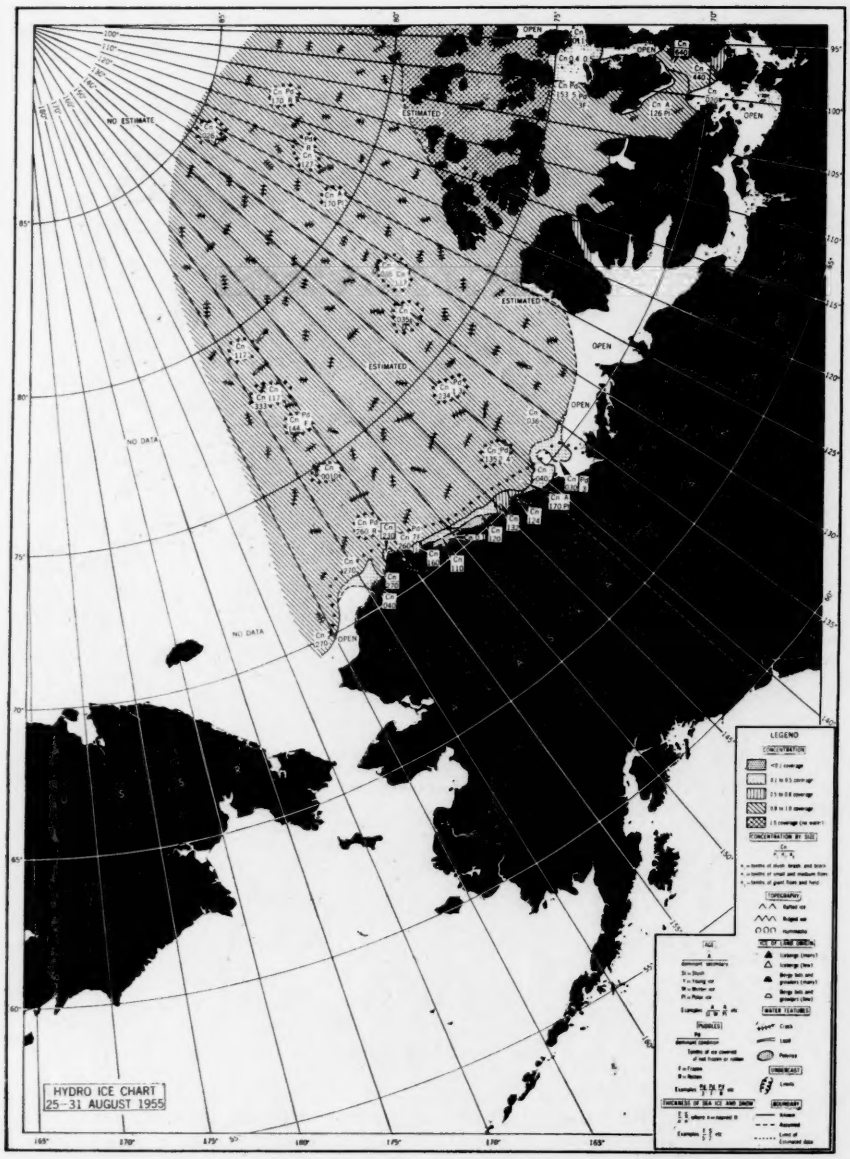


Fig. 7. Observed ice conditions 25–31 August 1955 off northern Alaska

1955 the ice moved onshore under heavy pressure throughout this distance, causing ships in the Point Barrow area to evacuate (see *Fig. 7*). Repeated onshore ice movements combined with converging ice-bearing currents at the Point proper kept all types of vessels from rounding Point Barrow in either direction from 25 August to 7 September 1955. Because of large amounts of ice in the Barrow unloading area to the south of the Point, only a few of the ships scheduled here were completely offloaded by the end of August 1955. The mid-August and early September 1955 boundaries of the polar pack shown in *Figs. 8* and *9* indicate this situation to have been the severest and most persistent of the 12 years the Navy had been hauling supplies to the Point Barrow establishment. The mean air temperatures (see *Fig. 2*) also reflected this anomalous condition, Wainwright and Point Barrow running 6°F and Barter Island 4°F below August normals.

The mean 5-day 700-mb map for 3–7 September 1955 (see *Fig. 10*) shows the polar vortex to have weakened considerably. Finally, between 5 and 8 September 1955 a 975-mb low-pressure area from Siberia stagnated in the Bering Sea (see *Fig. 11*). This storm, termed "the best damned storm I ever did see" by Vice-Admiral Denebrink, U.S.N., Commander of the Military Sea Transport Service, provided the long-awaited offshore winds and caused the polar pack to begin to recede finally. By 16 September 1955 the last elements of the Task Group completed their supply mission, cleared Point Barrow, and headed, according to their message, for a large polynya

Suppl.
Part. I
1956

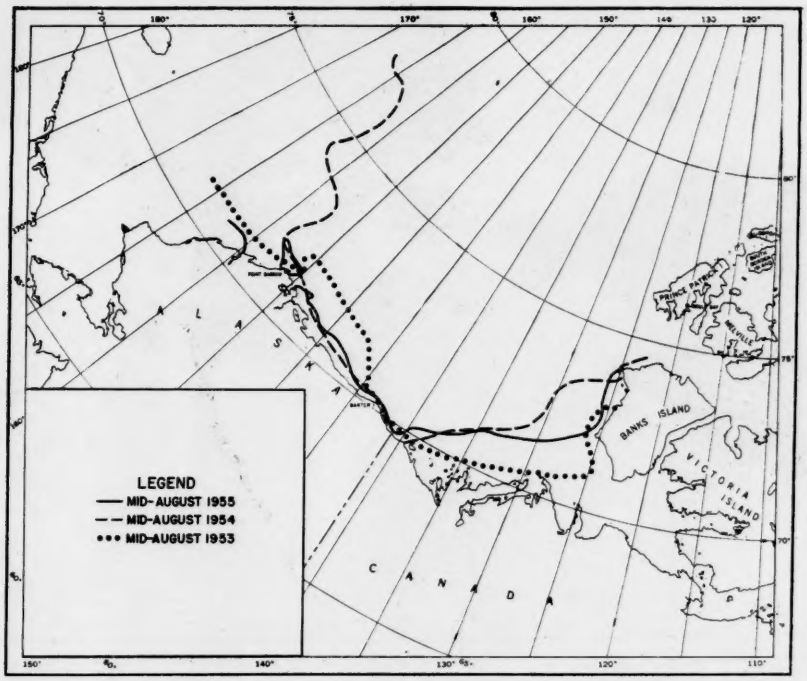


Fig. 8. Mid-August boundaries of the Polar Pack in 1953, 1954 and 1955 off northern Alaska

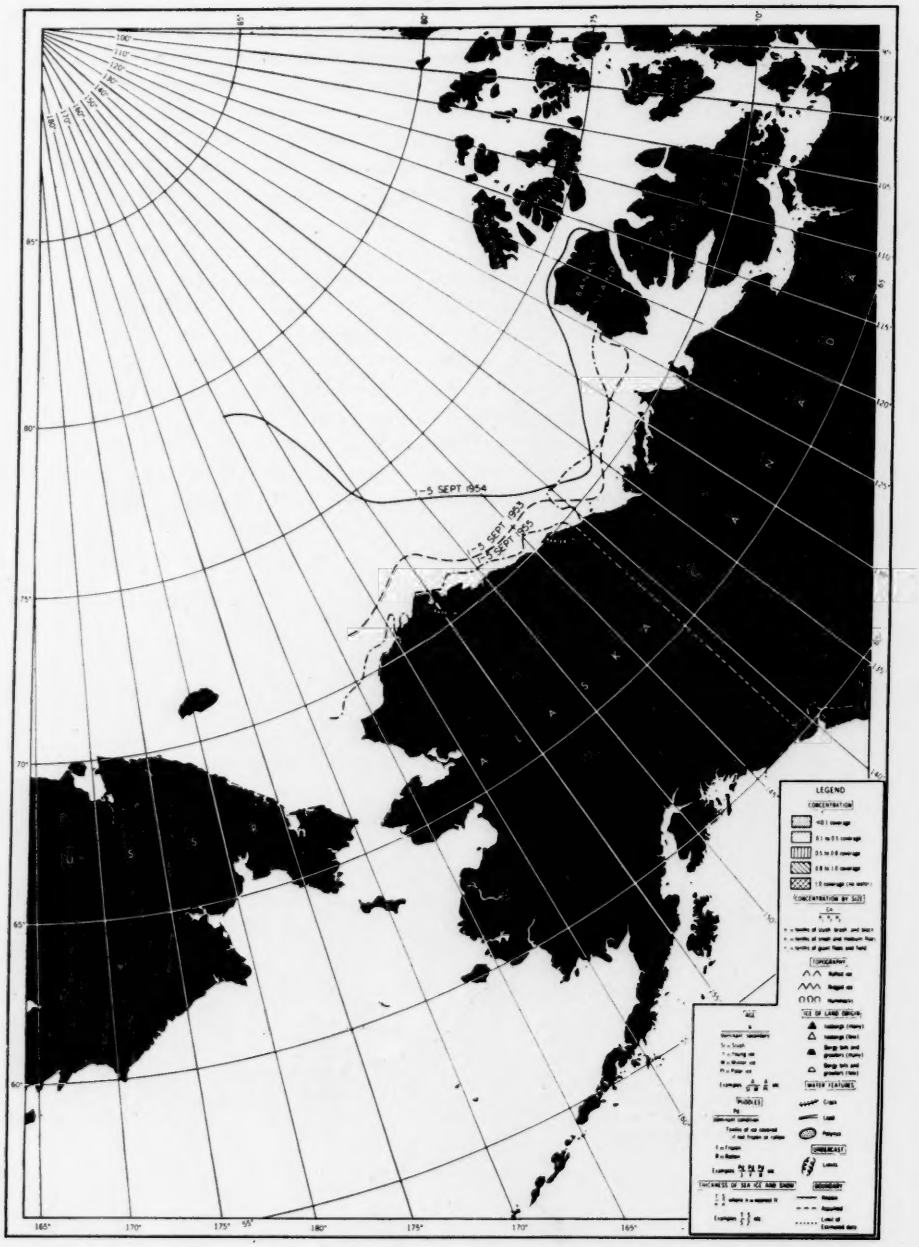


Fig. 9. Early September boundaries of the Polar Pack in 1953, 1954 and 1955 off northern Alaska

known to exist with San Diego at the far end where much cube ice was anticipated. Although all cargo was delivered and no ships lost, the severity of the environmental conditions experienced can be better understood when realizing the Task Group reported that only 4 out of fifty-seven ships passing along the northern Alaskan coast suffered no damage from grounding or ice.[2]

From the above discussion, it may be concluded that the anomalous ice conditions along northern Alaska in the summer of 1955 were at least partially caused by storm tracks in August and early September and were associated with the highly-developed polar vortex. It is apparent that there is still a lack of sufficient oceanographic and meteorological data on which to base a detailed explanation as to the controls exerted by inflowing warm water, by the behavior of an ice-covered basin as vast as the Arctic Ocean

Suppl.
Part. I
1956

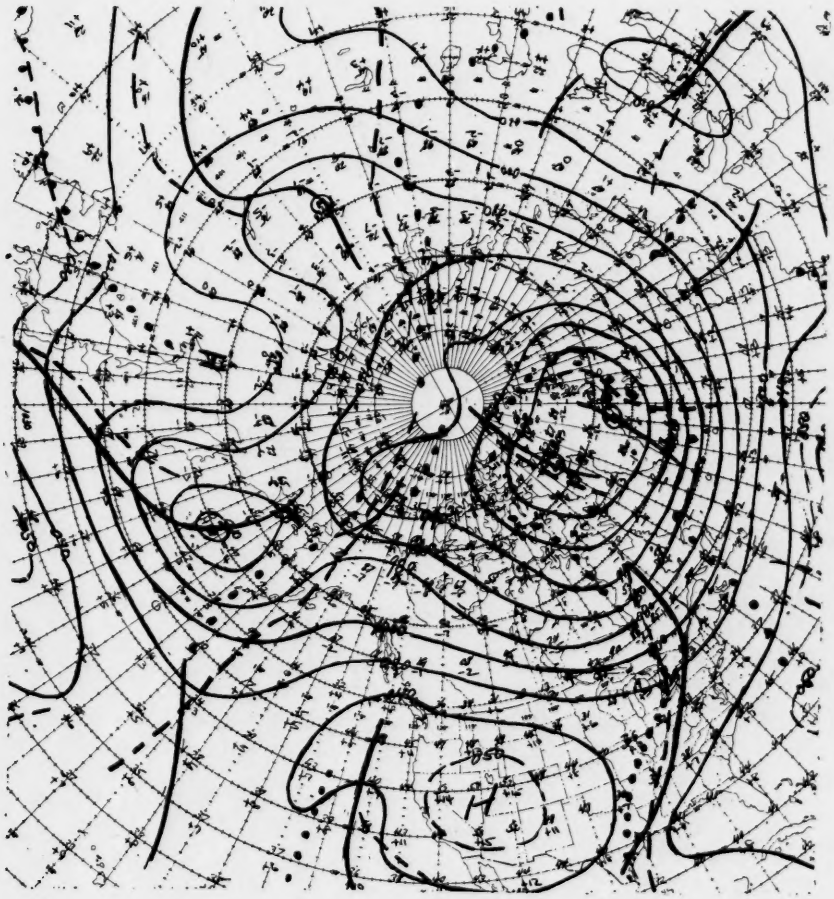


Fig. 10. Mean 5-day 700-mb level chart for 3–7 September 1955. (Courtesy U.S. Weather Bureau)

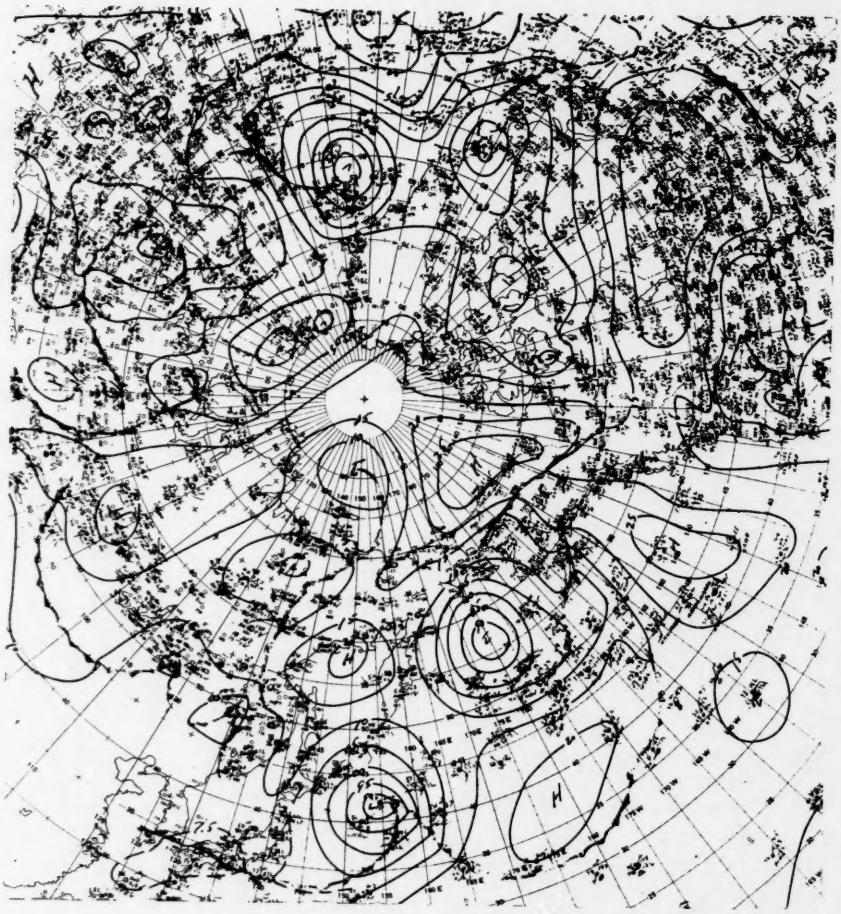


Fig. 11. Surface weather map for 1230Z, 8 September 1955. (Courtesy U.S. Weather Bureau)

under varying storm patterns, and by the change from a persistent warm spell to a persistent cold spell over Alaska in less than 30 days.

References

[1] Bates, C. C. "Sea Ice and Its Relation to Surface Supply Problems in the American Arctic" *Bull. geol. Soc. Amer.* Vol. 68, No. 12–2, p. 1711 (abstract), 1955

[2] Denebrink, F. C. "Cargo Ships Penetrate the Arctic" *Natl. Def. Transportation Jour.* Vol. 11, No. 6, pp. 30–34, 1955

DISCUSSION, 6 JULY 1956

Chairman: R. SCHERHAG, Germany

SVERDRUP: The remark I would make pertains to the paper by Dr. Fritz on the whiteout. It may be a matter of interest to say that it is a phenomenon quite well known in the Norwegian mountains above the timber-line, and that the mountain farmers in central Norway have a Norwegian word for it; they simply call it "snow darkness", directly translated. What occurs to me here is that possibilities for examination of the phenomenon ought indeed to be present here, where they might be carried out more readily and easily than at Arctic stations. I do hope some of my colleagues in Norway will be aware of that, and perhaps may be able to act.

FRITZ: With regard to the Arctic whiteout, I would very much like to see some work done on measurements of the distribution of sky brightness in visible light, at the same time as measurements of the surface albedo are made. We have sent an instrument to the Antarctic to try to do this. But I think that in the Antarctic we may find conditions where the surface albedo may not vary enough to enable us to say when we should have the whiteout and when we should not have it, in relation to the value of surface albedo. In other words if the albedo of the surface is always high we might not have enough varying conditions to set a lower limit of albedo during which the whiteout might exist. And in Norway probably you have these variations, and I think it will be very interesting to carry out a study here.

LILJEQUIST: In connection with Dr. Fritz's paper I should like to make some comments about the albedo and to refer to some measurements we made at Maudheim in the Antarctic.

We measured the albedo of the snow surface both with caloric methods, viz. with Moll-Gorczynski pyranometers, and with photoelectric (selenium) cells and Schott colour filters.

We then obtained with the photoelectric cells an albedo for visible light as high as 96 per cent. These measurements were made in overcast weather, so they should not be influenced by deviation of the instrument from the cosine law.

With caloric methods (giving "total" albedo) we obtained the albedo about 90 per cent in a dense overcast, and with a clear sky we got values around 80 per cent with the albedo increasing for decreasing solar latitude. Even from these data it is apparent that the albedo in the infrared must be decidedly lower than for visible light.

With a dense overcast the radiation flux penetrating the cloud will be reflected up and down between cloud-base and snow surface, and the sum of the different downward directed fluxes gives the observed global radiation.

In a very simplified manner we obtain from an infinite geometric series

$$\alpha = \frac{G}{G_o} = \frac{1-A_c}{1-A\,A_c}$$

where G and G_o is global radiation with clouds and with a clear sky respectively, A = snow-albedo, A_c = cloud-albedo.

Generally we can take $A_c = 75$ per cent. In temperate, not snow-covered regions $\alpha = 23$ per cent is generally found, while in the polar regions with extensive fields of (not melting) snow 60–70 per cent is observed. Of course, the density, drop-size, etc., of the clouds also affect α, but the main cause, at least for average

conditions, must be attributed to the multiple reflections between surface and cloud, i.e. ultimately to the albedo of the surface.

With $A = 100$ per cent we should obtain $\alpha = 100$ per cent from the simplified formula above, and with $A = 96$ per cent—as in our case for visible light—about 90 per cent is obtained. These are, of course, upper limits as no regard has been paid to absorption, etc., in the cloud.

The most remarkable is, however, the rapid increase of α with increasing surface albedo A, when the latter is high. For infrared solar radiation the albedo must be less than for visible light (compare our measurements previously mentioned), and as a consequence the global radiation with a dense overcast must be relatively more rich in visible light than with the sky clear. The night "total" albedo (about 90 per cent) in overcast weather must in part be attributed to this circumstance.

We also measured the angular distribution of the sky-brightness with a dense overcast. We used a photoelectric cell receiving light through a cone of aperture 15°. One measurement is given in *Fig. 1*. It shows a section facing directly inland and

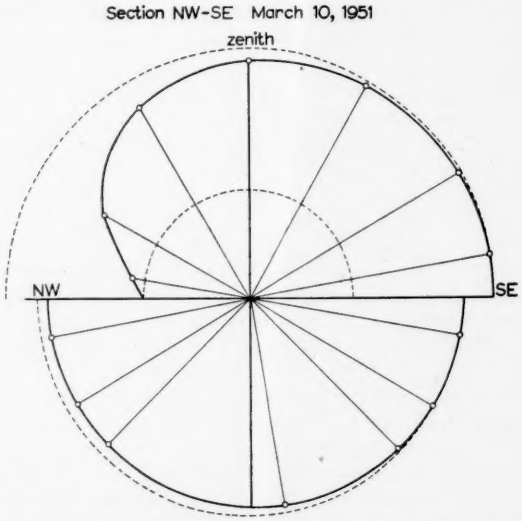


Fig. 1. Sky-brightness and intensity of the light reflected from the snow, dense overcast with "watersky" at Maudheim. The diagram shows the light-intensity in a section SE-NW, with open sea 5km away to NW and snow-covered inland to SE. (Dashed lines have been drawn to indicate the distribution of light-intensity more clearly.

directly towards the coast at 5 km distance and with open water beyond. Over the sea there was a marked "water-sky", i.e. here are practically no multiple reflections, but inland there are. In this latter direction the sky-brightness is for all practical purpose isotropic, which should be the case according to the theoretical considerations of Dr. Fritz. Also the reflected light from the snow is isotropic.

LANGLO: I should like to comment on some of the problems raised in the interesting paper by Harald Johansen. It may be pertinent to repeat here a remark made by Dr. Sutcliffe in an informal discussion this afternoon, namely, that in the early pioneer days of ozone research—when Dr. Dobson discovered the relation between the total amount of ozone and the position of high- and low-pressure areas at the earth's surface—the problems and their explanation seemed to be relatively simple. Now, when our knowledge of the ozone-weather relationship is much more detailed, the problems are becoming increasingly complicated.

To start with the observational data, particularly those illustrating the variation

of ozone with latitude, we do not know for certain the distribution of ozone at high latitudes during the winter months. There is some evidence, however, based on recent measurements with more sensitive instruments and on observations taken against the moon, that the earlier published values for Tromsö for the months November–January are too low and that the monthly mean ozone values for these months probably do not go below 0·200 cm.

If this is accepted as a working theory, we may start our consideration on the assumption that the ozone amount at a certain month—say October—is fairly uniformly distributed over the entire globe and having a magnitude of around 0·200 cm.

With reference to the suggestion made by H. Wexler that the extraordinarily large increase in the total ozone content observed in late winter at latitude 70°N may be explained by a breakdown of the strong zonal westerly winds in the lower stratosphere into a blocking pattern whose large meridional motions transport ozone from lower latitudes, I should like to mention an investigation on similar lines I made some years ago, but the results of which were not published. Based upon the values of the zonal index evaluated from the 500-mb charts for temperate latitudes (references: Namias and Breistein and Smebye) for each month of the period 1939–49, a comparison was made between these index values and the deviations observed in the monthly mean ozone values from the "normal" mean values based on the same 10-year period. No significant correlation was found in this way between the ozone variations and the variations in the zonal index for a pressure surface in the troposphere where a high correlation could be expected. One of the reasons for this negative result may be that the amount of ozone used in these correlations was measured in one point only and that the effect on the ozone distribution of the breaking down of the strong westerlies into a blocking pattern can only be more fully understood if we have a suitable network of ozone observing stations at high latitudes distributed both along a latitude and in meridional directions. It is important to bear in mind that the actual horizontal ozone gradients in late winter at around latitude 65°N are of the same order of magnitude in the direction north-south as in the direction east-west. The network of about 40 ozone observing stations planned to be established for the forthcoming International Geophysical Year will probably be sufficiently dense to throw further light on this fascinating subject.

It would be of particular interest for the meteorologists working in the field of long-range forecasting if a certain lag effect could be discovered between the changes in the total amount of ozone and other meteorological phenomena in the stratosphere or in the troposphere. Up to now, no such lag effects have been demonstrated.

SCHERHAG: In this connection I have a question. Are there any ozone measurements available within these stratospheric cold poles and within the stratospheric high-pressure centers? If there is any connection, then it must be found within these high-reaching systems. If we have measurements within these stratospheric systems, we shall get some information about the origin of the air-parcels. It is very difficult to draw any conclusion from advection only.

NAMIAS: I agree completely with Dr. Langlo regarding the necessity of considering the portion of the planetary wave with which one is dealing in order to fix the ozone in relation to the pattern. However, if we are able to get a network of ozone stations to arrive at a synoptic picture, so that we can make latitudinal and point averages, it is quite conceivable, and indeed likely, that there will be found connections between the form of the general circulation and the total ozone. In the paper represented by Johansen, the correlations between the pressure at various levels, and even in mid-troposphere, with ozone content were reasonably good and consistent. There are frequent occasions when the total hemispheric circulation partakes in an organized evolution. For example, a profile of the departure from normal of mid-tropospheric height over the Northern Hemisphere during the past winter of 1955–56 shows that

heights were much higher than normal in northern latitudes, north of about 40°N, with corresponding negative departures over lower latitudes. This large positive anomaly was associated with many weather factors, including the cold weather in Europe and the abnormal conditions in America and in the Pacific. Now, to the extent that the ozone distribution is correlated with these pressures, which seems rather likely, one could obtain a synoptic picture of large scale over the hemisphere. But, for a particular locality, the position with respect to the troughs and ridges must be considered.

It was discouraging to hear from Dr. Langlo that relationships found so far indicated only simultaneous and not lag interactions; that is, discouraging from the standpoint of possible utilization in weather prediction. Perhaps we may be able to predict ozone better than use it as a predictor.

MORELAND: In reply to Dr. Langlo, I want to emphasize that the zonal circulation which Dr. Wexler and I are considering is in the stratosphere and not in the troposphere. The 50-mb charts show that a rather large cold polar vortex maintains itself through the winter months. When the sun returns to higher latitudes in the spring the zonal patterns break down. The trough and ridge patterns during the spring have a much greater amplitude causing the import of ozone from lower latitudes. This advection combined with subsidence causes the sharp increase in ozone. An attempt was made to relate total amounts with the pressure patterns at 50 mb over north-west Europe. In the winter months, however, it was not feasible to do this, especially in January when the ozone data were so sparse that one could not construct daily ozone synoptic maps. However, during the spring months, the data were sufficient to construct daily ozone charts as well as charts showing the daily deviation of ozone from the monthly mean values. These deviation charts when compared with the 50-mb charts indicated that high ozone values are found in the north-east quadrant of a low, and the negative values are located in the center of the cold ridges at 50 mb. From these preliminary investigations there appears to be a rather strong relationship between pressure systems at 50 mb and the maximum and minimum amounts of ozone. If an areal synoptic chart of total ozone is superimposed on a 50-mb map, it is usually found that the ozone maxima and minima will delineate the troughs and ridges respectively, except that, as described earlier, the warm dynamic anticyclones which form at high latitudes are associated with a pronounced ozone maximum.

SVERDRUP: I would like to return to a question pertaining to radiation and make a remark which is perhaps somewhat naïve. The radiation phenomena which have been discussed here all deal with the characteristics of the surface. And it is obvious that part of the cooling of the stratosphere in high latitudes is directly related to the conditions at the surface. I had been hoping that comments might also have been made on the cooling of the free atmosphere of the polar regions, particularly during the season when the sun is below the horizon. There must exist a number of problems which are of somewhat different character from those encountered in lower latitude. In the Arctic there is no insolation during many months of the year and the cloudiness is small. Periods with clear sky prevail for considerable lengths of time, and whenever clouds are present they are much thinner than those which are found in lower latitudes. Furthermore, associated with the low temperature of the atmosphere, the water-vapour content is very low. Still, the atmosphere as such must loose large amounts of energy by radiation in the winter season. I am mentioning this because I would think it important to pay greater attention to the cooling of the free atmosphere. In this picture the characteristics of the subsurface doubtlessly also enters. The very low temperatures which the snow-covered surface attains must also be of importance to what goes on in the free atmosphere. I do not know if anybody working with radiation problems might now be willing to comment on these remarks.

GODSON: With regard to the free-air cooling, which arises as a result of infra-red radiative flux divergence, the most reliable calculations that I have seen for a continental Arctic air-mass give an almost constant rate of cooling throughout the layer from, say, 900 to 300 mb of about 0·8°C per day. If there are clouds present, the effect will be to enhance the radiative cooling above the cloud through a layer of about 100 mb; above that level the rate of cooling is very little affected. Below the cloud there will be a much reduced rate of cooling. There are still a number of uncertainties associated with such calculations, but it is likely that the correct answer is just under 1°C per day. With normal air-masses in temperate latitudes the cooling is probably of the order of 50 per cent greater, or even more.

SUTCLIFFE: May I ask Dr. Godson a question? Do calculations lead us to suppose that the very large changes in temperature in the stratosphere which are associated with ozone changes and so on, have any significant effect on tropospheric temperatures? In calculating troposphere temperature changes may we ignore stratospheric changes?

GODSON: It is probably quite safe to ignore specific stratospheric conditions in calculating radiative effects up to about 400 mb. Variations in the water-vapour content of the stratosphere would, however, significantly affect the cooling (or warming) at the tropopause. Thus, a relatively moist as well as warm stratosphere should lead to a radiative warming at the tropopause. Carbon-dioxide will give, of course, a net warming at the base of any temperature inversion. Water vapour will only give a net warming if its decrease in concentration above that level is not too great.

SUTCLIFFE: The question is in connection with a model of the general circulation of the atmosphere. Some authorities think the ozone is very important as controlling the temperature and the height of the tropopause. I was asking whether, if we were building up a model of the general circulation, and thinking of the winter cooling over the polar cap, we could safely ignore the ozone and these high-level phenomena.

GODSON: I believe it is possible to ignore the radiative contribution of ozone if one is interested in only the lowest 800 mb of the atmosphere. Even carbon-dioxide, which is a more effective absorbent than ozone by reason of both concentration and distribution of absorption bands through the spectrum, has a contribution to radiative changes which is small relative to that of water vapour. I believe it was formerly felt that ozone might be an important absorbent in the vicinity of the tropopause, but this view is no longer held.

NAMIAS: What Dr. Godson has been describing is a sort of average rate of cooling of the free atmosphere. But as we know, as Dr. Bates pointed out, the conditions in the polar region vary considerably from one year to the next, so that the resulting distribution of temperature in the lower troposphere varies considerably from one year to the next. The naïve question that I would like to ask Dr. Godson is how close are we to any system of incorporating these different factors into some sort of a numerical model. I mean is there any hope for including these variations from the normal cooling due to the state of the atmosphere in a given year.

GODSON: If one wished to include radiative temperature changes in a model for numerical prediction of tropospheric circulations, one would find that the effects of cloudiness on radiative cooling would predominate over the effects of variations of temperature or humidity on such cooling. In cases of clear skies, calculations with different typical air-masses, or with typical temperature and humidity distributions over a wide range of latitudes, give very similar results for the radiative cooling. However, the introduction of clouds can change these results appreciably. For example, the net cooling of the layer from 1000 to 500 mb will be quite different if a middle cloud deck lies entirely below 500 mb than if such cloud lies entirely above 500 mb, and both will differ from the case of clear skies. Compared to such variations, the effects of varying distributions of temperature and humidity with clear skies is relatively small.

Suppl.
Part. I
1956

LANGLO: I would like to draw attention to a paper by Dobson. He very clearly shows that because of the very dry stratosphere which has been confirmed by a number of ascents by aircrafts with dew-point hygrometer, it is fairly safe for temperate latitudes to say that humidity in the lower stratosphere is so low that the importance of the ozone is about of the same order with regard to the radiation effect. This is a very safe conclusion. He concludes that the effect of the ozone on the temperature of the lower stratosphere is of the same order as the effect of the water vapour present in the lower stratosphere.

SCHERHAG: I would like to draw attention to the strong temperature variations which occur at high levels over the Arctic, especially at 50, 30 or 20 mb. There are observed temperature changes I believe of more than 50°C within a few months. How should these be explained?

SVERDRUP: I would like to make a slight comparison between conditions in the Arctic and the Antarctic, at least as far as conditions at Maudheim are concerned. According to investigations by Liljequist the drifting snow started at Maudheim at wind velocities between 8 and 9 m/sec^{-1}, and one of the speakers, I believe Mr. Mitchell, mentioned a limit of 4 to 6 m/sec^{-1} for the Arctic. From our experience on board *Maud* the limit lies at about 4·5 m/sec^{-1}. The difference must be associated with the different character of the snow-cover. In the Antarctic the surface density of the snow-cover is high, evidently because of the very high wind velocity which packs the snow together solidly. The winds in the Arctic never reach velocities comparable to those observed in the Antarctic; the snow-cover is looser and the drift starts much earlier.

REED: I would like to ask some questions regarding Mr. Mitchell's paper. Has any attempt been made to correlate the haze with the synoptic situation? In other words, do we know whether it is more frequent under cyclonic or anticyclonic conditions? Perhaps the simplest thing to assume is that the haze is some sort of condensate and that it might therefore be related to the weather pattern.

Next, is the occurrence of haze related to temperature? Is it possible that it occurs only at very low temperatures so that it may be composed of very fine ice particles?

I am surprised that 12,000 ft is a representative height for cloud tops in winter. Does this mean that the clouds in question are composed at least partly of water droplets? At greater heights and lower temperature condensation may be in the form of very fine ice particles that only slightly reduce visibility.

MITCHELL: In attempting to answer Dr. Reed's first question, I am guided principally by the testimony of the reconnaissance observers who have flown on the "Ptarmigan" mission, and also by my own observation of the haze at the time I accompanied one of the missions myself. The haze characteristically occurs a high percentage of the time that the weather is otherwise clear. Generally, that means that there are no significant cloud layers present with the haze, except many times there may be a very low cloud deck over the ice of the Arctic Ocean. Dropsonde data indicate that quite often there are inversions, apparently subsidence inversions, present in the atmosphere below flight level (500 mb) at the time the haze is observed. In general it seems likely that there is a certain amount of subsidence accompanying the haze, although I do not know that the evidence can be called conclusive.

With regard to the second question, the temperatures recorded in this haze have usually been between —30° and —35°C. If water is at all involved, its state would probably be ice, although it might be supercooled water considering the very small particle sizes evidently present. I believe that a physical meteorologist would tell us that it is possible to have supercooling down to those temperatures. However, to the best of my knowledge, there has never been a report of icing in this haze.

With regard to the third question: When I referred to 12,000 ft being a representative top to clouds in the Arctic, I was talking only of the *solid* cloud decks. Of

course, higher clouds do frequently exist, and a high percentage of these are composed of ice crystals.

SVERDRUP: I may perhaps add a little to the last question, because we also had the opportunity of observing haze at ground level, about which also Mr. Mitchell spoke. We found there that at temperatures below — 35°, haze was relatively frequent even out in the pack-ice. And it is not associated with any particular weather situation, because the temperature is very closely related to the wind velocity such that haze will only occur in calm weather. In these conditions, as soon as it starts blowing, the temperature will rise, you get a mixing and under those conditions haze will not appear any more. That is just the experience from the pack-ice north of Siberia. I might add that the occurrence of those low temperatures agrees well with some of the graphs which were shown by Mr. Mitchell.

MITCHELL: I would like to say one further thing about this haze which is in the nature of speculation, I suppose. The question was raised as to what extent the haze is composed of water. I would estimate that the extent to which it is composed of water, as distinguished from lithometeoric material, will be about comparable to the extent to which haze in the lower latitudes is composed of water. And this, as I understand it, is not overwhelmingly the greater component. That means, then, that the presence of the haze might indicate some kind of a sink in the atmosphere, with respect to the general circulation, whereby a part of the dust and so forth carried into the atmosphere in lower latitudes might reach the polar regions.

FRITZ: It seems to me that we now get into the realm of the general circulation where we can possibly check some of these things. We could collect a mass of air in an airplane and by simple Owen's counters, for instance, condense out all the particles and examine them under a microscope, to determine their character and size. This sampling technique, which has been done many times, might help to check the suggestion made.

The visibility which you mentioned (something like 2–4 miles) are estimates. It is difficult to estimate the visibility from an airplane when the visibility is much greater than the length of the wingspan. And there again I would suggest a simple experiment which has been performed many times: namely, putting a light beam on the wing and a detector on the tail surface, to check what the actual transmission is under those conditions. The transmission may give an estimate of the visibility and can be compared with the transmission in haze-free air.

SUTCLIFFE: Has Dr. Mitchell obtained reliable measurements of humidity above the haze?

MITCHELL: No. The measurements in this aircraft have been made with wet- and dry-bulb thermometers. I believe that newer techniques of humidity measurement are being introduced at the present time. These older methods are little more reliable than the traditional surface measurements of humidity at very low temperatures. As a general matter, however, the humidities are fairly high, and one of the observers remarked to me that he thought that contrail formation was slightly more probable in the presence of this haze than otherwise. This would indicate higher humidities in the haze.

LANGLO: May I ask Mr. Schumacher whether any further measurements were made of the temperature inside the screen beside the normal instrument, for instance ventilation by Assmann inside the screen at the same time as outside? Can you find the reason for this difference in temperature? Can you find any reasonable explanation of a radiational effect?

SCHUMACHER: What I have supposed is that the walls are heated due to high intensity of the short-wave radiation, and so the bulb on the thermometer is heated by direct radiation from the wall of the screen. As to the temperature measurements, we did no Assmann readings in the screen. Beside the normal thermometer we had a standard thermograph.

Suppl. Part. I 1956